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Magma Plumbing Systems: A Geophysical Perspective

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Abstract

28 Over the last few decades, significant advances in using geophysical techniques to image the
29 structure of magma plumbing systems have enabled the identification of zones of melt
30 accumulation, crystal mush development, and magma migration. Combining advanced geophysical
31 observations with petrological and geochemical data has arguably revolutionised our understanding
32 of and afforded exciting new insights into the development of entire magma plumbing systems.
33 However, divisions between the scales and physical settings over which these geophysical,
34 petrological, and geochemical methods are applied still remain. To characterise some of these
35 differences and promote the benefits of further integration between these methodologies, we
36 provide a review of geophysical techniques and discuss how they can be utilised to provide a
37 structural context for and place physical limits on the chemical evolution of magma plumbing
38 systems. For example, we examine how Interferometric Synthetic Aperture Radar (InSAR), coupled
39 with Global Positioning System (GPS) and Global Navigation Satellite System (GNSS) data, and
40 seismicity may be used to track magma migration in near real-time. We also discuss how seismic
41 imaging, gravimetry, and electromagnetic data can image contemporary melt zones, magma
42 reservoirs, and/or crystal mushes. These techniques complement seismic reflection data and rock
43 magnetic analyses that delimit the structure and emplacement of ancient magma plumbing systems.
44 For each of these techniques, with the addition of full-waveform inversion (FWI), the use of
45 Unmanned Aerial Vehicles (UAVs), and the integration of geophysics with numerical modelling,
46 we discuss potential future directions. We show that approaching problems concerning magma
47 plumbing systems from an integrated petrological, geochemical, and geophysical perspective will
48 undoubtedly yield important scientific advances, providing exciting future opportunities for the
49 volcanological community.

50

51 **1. Introduction**

52 Igneous petrology and geochemistry are concerned with the chemical and physical mechanisms
53 governing melt genesis, mobilisation, and segregation, as well as the transport/ascent, storage,

54 evolution, and eruption of magma. The reasons for studying these fundamental processes include
55 understanding volcanic eruptions, modelling the mechanical development of magma conduits and
56 reservoirs, finding magma-related economic ore deposits, exploring for active geothermal energy
57 sources, and determining the impact of magmatism in different plate tectonic settings on the
58 evolution of the lithosphere and crustal growth. However, whilst petrological and geochemical
59 studies over the last century have shaped our understanding of the physical and chemical evolution
60 of magma plumbing systems, assessing the distribution, movement, and accumulation of magma in
61 the Earth's crust from these data remains challenging. A key frontier in igneous petrological and
62 geochemical research thus involves deciphering how and where magma forms, the routes it takes
63 toward the Earth's surface, and where exactly it is stored.

64 This contribution will demonstrate how geophysical data can be used to determine the
65 architecture of magma plumbing systems, providing a structural framework for the interpretation of
66 petrological and geochemical data. To aid the alignment of petrological, geochemical, and
67 geophysical disciplines it is first important to delineate what we mean by 'magma'. We follow
68 Glazner *et al.*, (2016) and define magma as, "naturally occurring, fully or partially molten rock
69 material generated within a planetary body, consisting of melt with or without crystals and gas
70 bubbles and containing a high enough proportion of melt to be capable of intrusion and extrusion".
71 Importantly, this definition specifically considers that magma: (i) forms through the migration and
72 accumulation of partial melt that is initially distributed throughout pore spaces in a rock volume;
73 and (ii) is a suspension of particles (i.e. crystals, xenoliths, and/or bubbles) within melt (see
74 Cashman *et al.*, 2017). As magma starts to solidify, the proportion of suspended crystals and thus
75 the relative viscosity of the magma increases until a relatively immobile, continuous network of
76 crystals and interstitial melt develops; we term this a 'crystal mush' (e.g., Hildreth, 2004; Glazner *et*
77 *al.*, 2016; Cashman *et al.*, 2017). The rheological transition from a magma to a crystal mush is
78 partly dependent on its chemistry, but typically occurs abruptly when the particle volume increases
79 across the 50–65% range (Cashman *et al.*, 2017). Crystal mushes thus exist at or above the solidus

80 and generally cannot be erupted, although they may be partly entrained in eruptible magma as
81 glomerocrysts, cumulate nodules, or restite (Cashman *et al.*, 2017). Migration of interstitial melt
82 within a crystal mush can lead to its accumulation and, thus, formation of a magma. A magma
83 plumbing system therefore consists of interconnected magma conduits and reservoirs, which store
84 magma as it evolves into a crystal mush, ultimately fed from a zone of partial melting (e.g., Fig. 1).
85 These definitions are supported by geophysical imaging and analyses of contemporary reservoirs,
86 which show melt volumes in the mid- to upper crust are typically low (<10%) and likely exist
87 within a crystal mush (e.g., Paulatto *et al.*, 2010; Koulakov *et al.*, 2013; Ward *et al.*, 2013;
88 Hammond, 2014; Comeau *et al.*, 2015; Comeau *et al.*, 2016; Delph *et al.*, 2017). These definitions
89 and geophysical data question the traditional view that magma resides in long-lived, liquid-rich, and
90 volumetrically significant magma chambers. Following this, the emerging paradigm for igneous
91 systems is thus that liquid-rich magma chambers are short-lived, transient phenomena with: (i) melt
92 typically residing in mushes that develop through the incremental injection of small, distinct magma
93 batches; and (ii) magma accumulating in thin lenses (e.g., Hildreth, 2004, Annen *et al.*, 2006;
94 Annen, 2011; Miller *et al.*, 2011; Solano *et al.*, 2012; Cashman & Sparks, 2013; Annen *et al.*, 2015;
95 Cashman *et al.*, 2017). We are now starting to view magmatic systems as a vertically extensive,
96 transcrustal, interconnected networks of magma conduits and magma/mush reservoirs (Fig. 1) (e.g.,
97 Cashman *et al.*, 2017).

98 The current use of geophysical techniques within the igneous community can be separated
99 into two distinct areas focused on either characterising active volcanic domains or investigating the
100 structure and emplacement of ancient magma plumbing systems. For example, in areas of active
101 volcanism, our understanding of magma plumbing system structure principally comes from the
102 application of geophysical techniques that detect sites of magma movement or accumulation (e.g.,
103 Sparks *et al.*, 2012; Cashman & Sparks, 2013). Such geophysical techniques include Interferometric
104 Synthetic Aperture Radar (InSAR; e.g., Biggs *et al.*, 2014), seismicity (e.g., recording of
105 earthquakes associated with magma movement; e.g., White & McCausland, 2016), various seismic

106 imaging methods (e.g., Paulatto *et al.*, 2010; Hammond, 2014), gravimetry (e.g., Battaglia *et al.*,
107 1999; Rymer *et al.*, 2005), and electromagnetic techniques (Desissa *et al.*, 2013; Comeau *et al.*,
108 2015). These techniques allow examination of: (i) the temporal development of magma plumbing
109 systems (e.g., Pritchard & Simons, 2004; Sigmundsson *et al.*, 2010); (ii) vertical and lateral
110 movements of magma (e.g., Keir *et al.*, 2009; Jay *et al.*, 2014); (iii) the relationship between
111 eruption dynamics, volcano deformation, and intrusion (e.g., Sigmundsson *et al.*, 2010;
112 Sigmundsson *et al.*, 2015); and (iv) estimates of melt sources and melt fractions (e.g., Desissa *et al.*,
113 2013; Johnson *et al.*, 2016). However, inversion of these geophysical data typically results in non-
114 unique, relatively low-resolution models of subsurface structures. Furthermore, some methods only
115 capture active processes, which may be short-lived or even instantaneous, potentially providing
116 information on only a small fraction of the magma plumbing system.

117 In contrast to the study of active volcanic domains, the analysis of ancient plumbing systems
118 through field observations, geophysical imaging techniques (e.g., reflection seismology, gravity,
119 and magnetic data), and/or rock magnetic experiments can provide critical insights into magma
120 emplacement, mush evolution, and allow the geometry of entire plumbing systems to be
121 reconstructed (e.g., Cartwright & Hansen, 2006; Stevenson *et al.*, 2007a; Petronis *et al.*, 2013;
122 Muirhead *et al.*, 2014; O'Driscoll *et al.*, 2015; Magee *et al.*, 2016). Whilst such studies of ancient
123 plumbing systems provide a framework for interpreting the structure of active intrusion networks,
124 capturing a snapshot of how magma moved and melt was distributed through the system at any one
125 time is difficult because magmatism has long since ceased.

126 All the techniques employed to define active and ancient plumbing systems, including
127 petrological and chemical analyses, provide information at different spatial and/or temporal
128 resolutions. Answering the major outstanding questions in studies of magma plumbing systems
129 therefore requires the integration of complementary petrological, geochemical, geophysical,
130 geochronological, and structural techniques. Here, we examine active plumbing systems using
131 InSAR, seismicity, seismic imaging, gravimetry, and electromagnetic techniques. To provide a

132 context for the interpretation of data pertaining to the active systems, we also discuss how seismic
133 reflection data and rock magnetic techniques can be used to derive the structure and evolution of
134 ancient intrusion networks. The potential of emerging techniques involving seismic full-waveform
135 inversion (FWI) and unmanned aerial vehicles (UAVs) are also considered, as is the role of
136 numerical modelling in bringing together outputs from different datasets. For each technique
137 described, we briefly discuss the methodology and limitations and provide a summary of the key
138 findings and potential uses, with a focus on integration with petrological and geochemical data. The
139 aim of this review is to facilitate and promote integration between petrologists, geochemists,
140 geochronologists, structural geologists, and geophysicists interested in addressing outstanding
141 problems in studies of magma plumbing systems.

142

143 **2. Understanding magma plumbing system structure**

144 Here, we discuss a range of techniques that can be utilised to define different aspects of magma
145 plumbing system structure and evolution. In particular, we describe how InSAR, seismicity, seismic
146 imaging (e.g., seismic tomography), gravity, and electromagnetic data is used to determine melt
147 fractions and distribution, track movement of magma in near real-time, and/or locate sites and
148 examine the evolution of magma/mush storage. Overall, these geophysical techniques allow the
149 structure of active plumbing systems and their transient evolution to be assessed. We also discuss
150 how seismic reflection data can provide unprecedented images of ancient plumbing systems and
151 associated host rock deformation in three-dimensions at resolutions of 10's of metres. Finally, we
152 examine the application of rock magnetic techniques to assess magma flow and crystallisation
153 processes at a range of scales.

154 Although beyond the scope of this review, it is critical to highlight that interpreting the
155 geophysical response of a rock or magma relies on understanding its physical and chemical
156 properties (e.g., density, temperature, and melt fraction). Laboratory experiments testing how rock
157 or magma properties influence geophysically measured parameters (e.g., seismic velocities and

158 resistivity) thus provide context for interpreting magma plumbing system structure and evolution
159 from geophysical data (e.g., Gaillard, 2004; Pommier *et al.*, 2010; Pommier, 2014).

160

161 **2.1. Insights into magma plumbing systems from ground deformation data**

162 *Technique*

163 Changes in volume within magma plumbing systems can deform the host rock, potentially resulting
164 in displacement of the Earth's surface. Such displacements are a unique source of information for
165 volcanologists and can be modelled to estimate geodetic source depth and, to varying extents, the
166 source geometry and volume change (e.g., Segall, 2010). Measuring the deformation of the Earth's
167 surface can thus provide information about the characteristics and timing of magma movement and
168 accumulation, as well as variations in internal reservoir conditions. Traditionally, deformation
169 measurements are made using levelling, electronic distance meters, tiltmeters, and Global
170 Positioning System (GPS), all of which have proven to be reliable methods and thus are widely
171 used in volcano monitoring (e.g., Dzurisin, 2006). For example, GPS measurements retrieve the
172 relative positions of receivers on Earth's surface from dual frequency carrier phase signals
173 transmitted from GPS or Global Navigation Satellite System (GNSS) satellites with precisely
174 known orbits. Distances between satellites and receivers are assessed from the travel-time, i.e. the
175 measured difference between the transmitted and received times of a unique ranging code, allowing
176 movement of the Earth's surface over time to be monitored (see review by Dixon, 1991).
177 Permanently installed receivers record position data continuously, but receivers can also be
178 deployed for a limited time during GPS campaigns to provide additional measurements, normally
179 made relative to a standard benchmark location (e.g., Dvorak & Dzurisin, 1997). Whilst tiltmeters
180 and GPS can provide continuous measurements, their spatial resolution is limited by logistical
181 constraints such as cost and accessibility, which may be restricted at active volcanoes.

182 The geographic reach of volcano geodesy has been greatly expanded over the past two
183 decades by the application of Interferometric Synthetic Aperture Radar (InSAR), an active remote

184 sensing technique that uses microwave electromagnetic radiation to image the Earth's surface (e.g.,
185 Simons & Rosen, 2007; Pinel *et al.*, 2014). Surface displacements can be measured by constructing
186 interferograms, where the difference in phase between radar echoes from time-separated images
187 appear as 'fringes' of variation in the line of sight distance to the satellite (Fig. 2). The patterns of
188 fringes in individual interferograms are distinctive for different deformation source geometries,
189 such as for horizontal (sill-like) or vertical (dyke-like) opening of intrusions, or the pressurisation of
190 a spheroidal reservoir (i.e. a Mogi source) (e.g., Fig. 2b). However, magma intrusion processes can
191 rarely be uniquely identified from geodetic source geometry alone, and distinguishing between
192 magmatic, hydrothermal, structural (e.g., faulting and compaction), and combinations of elastic and
193 inelastic sources is particularly challenging (e.g., Galland, 2012; Holohan *et al.*, 2017).

194 Whilst a single interferogram only provides displacements in satellite line-of-sight, a
195 pseudo-3D displacement field can be estimated by combining multiple images from polar orbits that
196 are ascending (i.e. satellite moves roughly northward, looking east) and descending (i.e. satellite
197 moves roughly southward, looking west) (Fig. 2a), especially where GNSS measurements can also
198 be incorporated. The lateral spatial resolution of most InSAR data is on the order of metres to tens
199 of metres, whilst vertical movements can be resolved on the order of centimetres and sometimes
200 millimetres. Temporal resolution depends on the satellite revisit time and ranges between days to
201 months depending upon the sensor type and satellite orbit. This means that InSAR can be used to
202 regularly assess ground deformation at virtually any volcano worldwide situated above sea level,
203 with a higher spatial density of measurements than achieved using from ground-based
204 instrumentation.

205 Magmatic processes are only observable by InSAR when either magma movement or
206 internal reservoir processes (e.g., cooling and contraction, phase changes) cause changes in pressure
207 and thereby instigate deformation of the host rock and free surface. The best-fit parameters of a
208 deformation source (e.g., an intruding magma body) are most often assessed by inverting measured
209 displacements using analytical elastic-half space models of simple source geometries, although

210 there are often trade-offs between parameters such as source depth and volume change (e.g.,
211 Pritchard & Simons, 2004). Complex and more realistic deformation source geometries may be
212 retrieved using finite element-based linear inversion of displacement fields (e.g., Ronchin *et al.*,
213 2017). A proportion of any pressure change may be accommodated by magma compressibility,
214 leading to underestimation of volume changes (e.g., Rivalta & Segall, 2008; McCormick-Kilbride *et*
215 *al.*, 2016). Assessing both volume changes and especially the total volume of a magma reservoir
216 from geodetic data therefore remains challenging. Furthermore, host rocks in areas of repeated
217 intrusion that have been heated above the brittle-ductile transition are better described by a
218 viscoelastic rheology (e.g., Newman *et al.*, 2006; Yamasaki *et al.*, 2018), while ductile
219 accommodation of volume changes may occur at greater depth. Where some constraints are
220 available for the structure and rheology of Earth's crust, finite or boundary element models may
221 achieve a more realistic model of the deformation source (e.g., Masterlark, 2007; Hickey *et al.*,
222 2017; Gottsmann *et al.*, 2017).

223

224 ***Observations***

225 Measurements of volcano deformation preceding and/or accompanying eruption have provided
226 insights into the extent and structure of magma plumbing systems and, in some instances, the
227 dynamics of magma movement. For example, InSAR-based observations at Eyjafjallajökull,
228 Iceland have recognised the intrusion of multiple, distinct sills over a decade and their subsequent
229 extraction when tapped during an explosive eruption (e.g., Pedersen & Sigmundsson, 2006;
230 Sigmundsson *et al.*, 2010). Extensive lateral connections via dykes and sills between reservoirs
231 and/or volcanoes have been illuminated by eruptions or unrest accompanied by ground deformation
232 tens of kilometres away, and by the existence of multiple deformation sources (e.g., Alu-Dalafilla
233 shown in Figures 3 and b, Pagli *et al.*, 2012; Korovin, Lu & Dzurisin, 2014; Cordon-Caulle, Jay *et*
234 *al.*, 2014; Kenyan volcanoes, Biggs *et al.*, 2014; global synthesis, Ebmeier *et al.*, 2018). Inter-
235 eruptive deformation at calderas is especially complex and seems to be particularly frequent and

236 high magnitude (e.g., Laguna del Maule; Fournier *et al.*, 2010; Singer *et al.*, 2014; Le Mével *et al.*,
237 2015), with the location of the deformation sources inferred to vary over time (e.g., Campi Flegrei,
238 Trasatti *et al.*, 2004; Yellowstone, Wicks *et al.*, 2006). The geometries of dykes and sills inferred
239 from InSAR data inform our understanding of changing subsurface stress fields (e.g., Afar,
240 Hamling *et al.*, 2010; Fernandina, Bagnardi *et al.*, 2013), as do measurements of displacements
241 caused by moderate earthquakes in close proximity to magma plumbing systems (e.g., Kilauea,
242 Wauthier *et al.*, 2013; Chiles-Cerro Negro, Ebmeier *et al.*, 2016).

243 At a transcrustal scale, deformation measurements have contributed to evidence for temporal
244 variations in magma supply rates (e.g., in Hawaii, Poland *et al.*, 2012). Volume increases in the
245 mid- to lower-crust, notably in the Central Andes, have provided the first observations of deep
246 pluton growth (Pritchard & Simons, 2004). Furthermore, uplift during episodes of unrest that have
247 not (yet) resulted in eruption have been detected at a broad range of volcanoes (e.g., Westdahl,
248 Mount Peulik, Lu & Dzurisin, 2014; Alutu and Corbetti, Biggs *et al.*, 2011) and, in some cases,
249 have been interpreted as evidence for the ‘pulsed’ accumulation of potentially eruptible magma
250 (e.g., Santorini, Parks *et al.*, 2012). In addition to magma movement, volume changes associated
251 with internal reservoir processes can also cause deformation of the host rock and free surface. For
252 example, InSAR measurements have recorded subsidence linked to cooling and crystallisation of
253 sills (Medicine Lake, Parker, 2016; Taupo Volcanic Zone, Hamling *et al.*, 2015). Transient periods
254 of subsidence during inter-eruptive uplift have been attributed to phase transitions in response to the
255 addition of more juvenile magma (e.g., Okmok, Caricchi *et al.*, 2014).

256

257 ***Implications and integration***

258 InSAR has increased the number of volcanoes where measurements of ground deformation have
259 been made, from less than 50 in the late 1990s to over 200 today (Biggs & Pritchard, 2017; Ebmeier
260 *et al.*, 2018). This increase in coverage has been particularly influential in the developing world
261 where monitoring infrastructure is typically poor (Ebmeier *et al.*, 2013; Chaussard *et al.*, 2013),

with InSAR often providing the first evidence of magmatic activity at many volcanoes previously considered to be inactive (e.g., Pritchard & Simons, 2004; Biggs *et al.*, 2009; Biggs *et al.*, 2011; Lu & Dzurisin, 2014). A continued increase in the number and range of satellite- and large-scale UAV-based SAR instruments, as well as enhancements to their spatial and temporal resolution, will allow the detection of a greater range of volcanic ground deformation (e.g., Salzer *et al.*, 2014; Schaefer *et al.*, 2015; Stephens *et al.*, 2017). Overall, improved InSAR coverage will also increase the number of volcanoes where deformation measurements have been made across multiple cycles of eruption and deformation, increasing its usefulness for both hazard assessment and for characterising the extent, geometry, and changes in magma plumbing systems.

Geodetic measurements provide information only about the parts of a plumbing system that are currently active, and do not necessarily reflect the full extent and character of the intrusion network (e.g., Sigmundsson, 2016). However, geodetic analyses of ground deformation provide critical insight into the spatial and temporal development of active plumbing systems. Comparing observations of ancient plumbing systems (e.g., Magee *et al.*, 2013; Schofield *et al.*, 2014), integration of ground deformation measurements with petrological observations (e.g., Caricchi *et al.*, 2014; Jay *et al.*, 2014) or thermal models (Parker *et al.*, 2016), as well as tomographic geophysical imaging, will increase the sophistication of models of magmatic systems. Integrating InSAR with gravity or electromagnetic measurements is particularly powerful, as it can allow discrimination between melt, volatiles, and hydrothermal fluids for which deformation signals are similar (see section 2.4) (e.g., Tizzani *et al.*, 2009).

282

2.2. Seismicity and magma plumbing systems

Technique

Seismicity (i.e. earthquakes) at volcanoes is primarily caused by the dynamic interaction of magma and hydrothermal fluids with the solid host rock (e.g., Chouet & Matoza, 2013), as well as by fracturing and fragmentation of silicic magma (e.g., Tuffen *et al.*, 2008). There are a number of

288 primary physical mechanisms for causing volcano seismicity (e.g., faulting), each of which
289 typically produces seismic signals of specific frequency content (Chouet & Matoza, 2013).
290 Recording and isolating different volcano seismicity signals therefore allows a variety of plumbing
291 system processes to be assessed. The majority of volcano monitoring agencies have now deployed
292 or aim to use a network of distributed seismic sensors, including broadband seismometers, to
293 monitor volcano activity (Neuberg *et al.*, 1998; Sparks *et al.*, 2012). Furthermore, an increase in
294 computing power and reduction in cost of seismic sensors means that researchers are now
295 developing fast, fully automated detection and real-time location techniques that can locate
296 seismicity to sub-decimetre precision (e.g., Drew *et al.*, 2013; Sigmundsson *et al.*, 2015).

297

298 ***Observations***

299 Volcano-tectonic (VT) seismicity generally produces relatively high frequency (1–20 Hz), short
300 period signals, involving clear primary (P), secondary (S), and surface waves, which are caused by
301 displacement on new or existing faults in the host rock in response to fluid-induced stress changes
302 (e.g., Rubin & Gillard, 1998; Roman & Cashman, 2006; Tolstoy *et al.*, 2008). These earthquakes
303 commonly occur near the propagating edge of intrusions, meaning the space-time evolution of VT
304 earthquake locations can be used to track the horizontal and vertical growth of sills and dykes (e.g.,
305 Keir *et al.*, 2009; Sigmundsson *et al.*, 2010; Sigmundsson *et al.*, 2015). Inflation of a magma or
306 mush body can also induce VT seismicity on any preferentially oriented faults surrounding the
307 intrusion, thereby recording the delivery time and locus of new magma injected into a reservoir
308 (e.g., Roman & Cashman, 2006; Vargas-Bracamontes & Neuberg, 2012).

309 Earthquakes with longer period seismic signals and low-frequencies (0.5–2 Hz) are thought
310 to be generated near the interface between magma and solid rock (Chouet & Matoza, 2013). The
311 earthquake source proximity to the magma causes the seismic signal to resonate in parts of the
312 plumbing system (e.g., conduits, dykes, and cracks), leading to a reduction in its frequency content
313 (Chouet & Matoza, 2013). These earthquakes can potentially be caused by stick-slip motion

314 between the magma and wall-rock or fracturing of cooling magma near the conduit wall (Neuberg
315 *et al.*, 2006; Tuffen *et al.*, 2008). Such earthquakes typically occur at restricted portions of conduits
316 where the magma flow and shear strain rate are highest (Neuberg *et al.*, 2006; Tuffen *et al.*, 2008).

317 Very long period seismicity (VLP) of 10s of seconds to several minutes period are typically
318 attributed to inertial forces associated with perturbations in the flow of magma and gases through
319 conduits (Chouet & Matoza, 2013). These signals can record the response of the host rock to
320 reservoir inflation and deflation and may be used to model conduit shape and size (Chouet *et al.*,
321 2008). To do this requires a better understanding of the links between flow processes and resultant
322 pressure/momentum changes using laboratory experiments and numerical models that include the
323 elastic response to magma flow across multiple signal frequency bands (e.g., Thomas & Neuberg,
324 2012).

325

326 ***Implications and integration***

327 Studies of evolving reservoirs now aim to link episodes of seismicity related to new magma
328 injection to petrological evidence for timing of reservoir recharge events, thereby providing
329 independent constraints on day to year-long time-scales of magma residence and input prior to
330 eruptions. For example, Fe-Mg diffusion chronometry modelling of orthopyroxene crystals from the
331 1980–1986 eruption of Mount St. Helens indicates that compositionally distinct rims grew within
332 12 months prior to eruption (Fig. 4) (Saunders *et al.*, 2012). Peaks in crystal growth correlated
333 extremely well with increased seismicity and SO₂ flux (Fig. 4), confirming the relationship between
334 seismicity and magma movement, as well as demonstrating how a combination of seismicity and
335 petrological information can be used to detect magma injections (Saunders *et al.*, 2012).

336 Petrology and seismicity can also be integrated with other methods, such as GPS and
337 InSAR. Field *et al.*, (2012) analysed volatiles in melt inclusions trapped in phenocrysts within
338 peralkaline lavas from historic eruptions at the Dabbahu Volcano in Afar, Ethiopia. Volatile
339 saturation pressures at typical magmatic temperatures were constrained to be in the range 43–207

MPa, consistent with the phenocryst assemblage being stable at 100–150 MPa. The interpreted magma/mush storage depths for these historic eruptions are ~1–5 km, consistent with the depths of earthquakes associated with reservoir inflation following dyke intrusion in 2005–2006 (Fig. 5) (Ebinger *et al.*, 2008; Field *et al.*, 2012). Additionally, the best-fit result for modelling of uplift patterns recorded by InSAR data, which were collected over the same time period as seismicity measurement, suggests the magma/mush reservoir comprises a series of stacked sills over a ~1–5 km depth range (Fig. 5) (Ebinger *et al.*, 2008). The consistency of depth estimates based on petrological study of ancient eruptions, along with the seismicity and inflation of the Dabbahu Volcano following axial dyke intrusion in 2005–2006, implies a vertically extensive and potentially long-lived magma/mush storage region. Such multidisciplinary studies demonstrate that joint observations and modelling of seismic signals, petrological data, and other techniques (e.g., geodesy and gas emissions) significantly strengthen interpretation of the physical structure, emplacement, and evolution of magma plumbing systems.

2.3. Identifying melt in plumbing systems using seismic imaging

Techniques

Both active and passive source seismological techniques, which utilise man-made seismic events and natural earthquakes respectively, can be used to identify areas where the presence of partial melt or magma causes a local reduction in seismic wavespeed, an increase in anisotropy, or an increase in attenuation (e.g., Berryman, 1980; Hammond & Humphreys, 2000a, b). With the recent availability of dense seismic networks, resolution of the crust and mantle seismic velocity structure has improved to the degree that active source seismic experiments can: (i) use tomographic techniques to image likely storage regions in the upper crust beneath ocean island volcanoes (e.g., Soufrière Hills Volcano, Montserrat; Fig. 6) (Paulatto *et al.*, 2010; Shalev *et al.*, 2010) and, occasionally, onshore volcanoes (e.g., Mt Erebus, Antarctica, Zandomenighi *et al.*, 2013; Mt. St. Helens, Kiser *et al.*, 2014); and (ii) utilise reflected data to image individual sills beneath mid-ocean

366 ridges (e.g., Kent *et al.*, 2000, Marjanovic *et al.*, 2014). A further example from Katla volcano
367 Iceland, demonstrates how active source seismic experiments can be used to identify S-wave
368 shadow zones (i.e. S-waves cannot travel through fluids) and delays in P-waves, which may be used
369 to infer the location and geometry of shallow-level magma reservoirs (Gudmundsson *et al.*, 1994).
370 However, recent modelling approaches suggest that the upper crust likely represents only a small
371 portion of magma plumbing systems and long-term storage is dominated by mushy zones
372 throughout the lower crust (e.g., Annen *et al.*, 2006). Active source seismic experiments,
373 particularly on land where the crust is thick and coverage less uniform, cannot penetrate to these
374 depths efficiently. Furthermore, whilst seismic tomographic methods using local earthquakes offer
375 3D images of crustal velocity beneath many volcanoes (e.g., Mt. St. Helens, Waite & Moran, 2009;
376 Askja, Iceland, Mitchell *et al.*, 2013), they can only resolve areas directly above the deepest
377 earthquakes. Non-uniform coverage thus makes interpreting tomographic images difficult as
378 resolution varies across the model (see review by Lees, 2007).

379 To illuminate lower crustal regions, seismologists rely on passive seismology. Extending
380 seismic tomographic images of magma plumbing systems to lower crustal depths requires the use of
381 teleseismic body-wave and surface wave data, which emanate far (>1000 km) from the
382 measurement site. However, these data are dominated by longer period signals, meaning their
383 resolution is relatively low. For example, the Fresnel zone (i.e. the region within $\frac{1}{4}$ seismic
384 wavelength and an estimate of the minimum resolvable structure) for active source data at 10 Hz is
385 on the order of 3 km in the upper crust compared to 10–15 km for 1 Hz teleseismic data used in
386 receiver function or tomography studies.

387

388 ***Observations***

389 Active and passive seismological techniques provide crucial insight into transcrustal melt and
390 magma distribution. For example, P-wave seismic travel-time tomography across Monserrat and the
391 Soufrière Hills Volcano images a series of relatively fast seismic velocity zones, which are

392 interpreted as solidified andesitic intrusions, surrounded by regions of slow seismic velocities likely
393 related to either areas of hydrothermal alteration or buried volcanoclastic deposits (Fig. 6) (Paulatto
394 *et al.*, 2010; Shalev *et al.*, 2010). Within the lower crust, inversions using surface wave data
395 generated by ambient seismic noise and receiver function data, which isolates P-wave to S-wave
396 conversions at major discontinuities in the earth, have identified low shear-wave velocities probably
397 related to melt presence beneath several volcanic settings (e.g., New Zealand, Bannister *et al.*, 2007;
398 Toba, Sumatra, Stankiewicz *et al.*, 2010; Ethiopia, Hammond *et al.*, 2011; Jaxybulatov *et al.*, 2014;
399 Costa Rica, Harmon & Rychert, 2015).

400 When trying to determine how much melt or magma is present, numerous studies have
401 shown that seismic velocities are more sensitive to the shapes of melt/magma-filled spaces on a
402 range of scales compared to the melt fraction (e.g., Hammond & Humphreys, 2000a, b; Miller &
403 Savage, 2001; Johnson & Poland, 2013; Hammond & Kendall, 2016). On the grain-scale, melt
404 commonly wets grain boundaries, forming planar pockets (e.g., Takei, 2002; Garapic *et al.*, 2013;
405 Miller *et al.*, 2014), whereas on the larger scale magma may form planar intrusions of either mush
406 (e.g., Annen *et al.*, 2006), or liquid-rich dykes or sills. If these features are preferentially aligned,
407 they will appear as a distributed region of melt to seismic waves and the analyses described will not
408 be able to discriminate between a melt-poor region dominated by aligned melt-pockets on grain
409 boundaries and an elongate melt-rich body such as an intrusion (e.g., Hammond & Kendall, 2016).
410 A further problem is that seismic velocities are affected by variations in temperature (Jackson *et al.*,
411 2002), composition (Karato & Jung, 1998), and attenuation (Goes *et al.*, 2012). Relating seismic
412 velocity anomalies to melt fraction is therefore difficult without some prior knowledge of melt
413 distribution (Hammond & Kendall, 2016).

414 One possible approach to investigate melt distributions further is through measuring seismic
415 anisotropy. If melt has some preferential distribution on a length-scale smaller than the seismic
416 wavelength, such as a stacked network of sills or an anisotropic permeability on the grain scale,
417 then the seismic wavespeed will vary with direction of propagation, i.e. be anisotropic. As a result,

418 measuring the effects of seismic anisotropy allows inferences about sub-seismic wavelength
419 structures, leading to estimates of the preferential orientation of melt distribution. It is common to
420 observe strong anisotropy beneath volcanoes and this has been used to place constraints on melt
421 distribution. For example, high degrees of shear-wave splitting from volcanic earthquakes can either
422 directly map out regions of significant quantities of melt aligned in pockets (Keir *et al.*, 2011), or
423 map out stress changes related to overpressure from injections of magma into the upper crust (Gerst
424 & Savage, 2004; Roman *et al.*, 2011). To image the deeper crustal magmatic system, azimuthal
425 variations in the ratio of P-wave to S-wave speeds (i.e. V_p/V_s) from receiver functions led to the
426 interpretation that a stacked network of sills is present in the lower crust beneath the Afar
427 Depression, Ethiopia (Hammond, 2014). Differences in the velocity of Rayleigh Waves and Love
428 Waves, which are vertically polarised shear-waves and horizontally polarised shear waves
429 respectively, suggest a similar anisotropic melt distribution is present beneath the Toba Caldera,
430 Sumatra (Jaxybulatov *et al.*, 2014) and Costa Rica (Harmon & Rychert, 2015).

431

432 ***Implications and integration***

433 Due to the interference of signals denoting the geometry of melt-filled pockets and the volumetric
434 proportion of that melt, estimating melt fraction remains difficult using seismology alone. Some
435 attempt has been made to directly infer magma/mush reservoir properties from seismic velocities.
436 For example, Paulatto *et al.*, (2012) used thermal modelling to test the range of melt fractions that
437 could account for the low velocity zones imaged in the upper crust beneath Soufrière Hills Volcano
438 (Fig. 6), Montserrat and concluded the melt fraction is between 3 and 10%. However, accounting
439 for resolution of the tomography, together with uncertainties in the distribution and geometry of
440 melt, means >30% melt may be present more locally in the low velocity zones defined beneath
441 Soufrière Hills Volcano (Paulatto *et al.*, 2012). Possible ways forward involve integrating
442 seismological data with: (i) petrological data that can place limits on likely melt fractions and/or
443 emplacement depths (e.g., McKenzie & O’Nions, 1991; Comeau *et al.*, 2016); (ii) geochemical

444 techniques that can help determine timescales of melt and magma evolution (e.g., Hawkesworth *et*
445 *al.*, 2000); and (iii) geodetic or other monitoring data, which helps determine magma movement
446 (Sturkell *et al.*, 2006). Recent efforts applying industry software, such as full waveform inversions
447 (FWI; Warner *et al.*, 2013), which is discussed in section 3.1, are also pushing the potential
448 application of seismological data further and mean that it may be possible to resolve features to sub-
449 kilometre levels, particularly in the upper crust. Together, these techniques may allow us to directly
450 relate seismic velocity anomalies to melt fractions and distributions in the whole crust.

451

452 **2.4. Studying magma plumbing systems using gravimetry**

453 ***Techniques***

454 Gravimetry measures the gravitational field and its changes over space and time, which can be
455 related to variations in the subsurface distribution and redistribution of mass (e.g., magma). A
456 variety of gravimeter instruments (e.g., free-fall, superconducting, and spring-based) and techniques
457 (e.g., ground-based, sea-floor, ship-borne, and air-borne instrumentations) are available. Spring
458 gravimeters, where a test mass is suspended on a spring, are mostly used to study magmatic and
459 volcanic processes in ground-based surveys (e.g., Carbone *et al.*, 2017; Van Camp *et al.*, 2017).
460 Changes in the gravitational acceleration across a survey area shorten or lengthen the spring, which
461 is recorded electronically and converted to gravity units. These changes are evaluated across a
462 survey network in relation to a reference and are hence termed ‘relative measurements’. Absolute
463 gravimetry can also be measured, i.e. the value of gravitational acceleration, and serves primarily to
464 create a reference frame into which other geodetic methods (e.g., InSAR, GNSS, levelling, relative
465 gravimetry) can be integrated for joint data evaluation. Recent reviews by Carbone *et al.*, (2017)
466 and Van Camp *et al.*, (2017) provide a broad account of gravimetric instruments, measurement
467 protocols, and data processing relevant for the study of magmatic systems.

468 Static gravimetric techniques obtain a single snap-shot of the subsurface mass distribution.

469 For example, Bouguer anomaly maps are perhaps the best-known products of static gravity surveys

470 and capture spatial variations in gravity over an area of interest, providing insight into anomalous
471 mass distribution in the subsurface. Within magmatic studies, computational modelling and
472 inversion of Bouger anomaly data allows identification of shallow intrusions (e.g., dykes and sills;
473 Rocchi *et al.*, 2007), magma-related ore bodies (Hammer, 1945; Bersi *et al.*, 2016), and plutons
474 (e.g., Figs 7a and b) (e.g., Vigneresse, 1995; Vigneresse *et al.*, 1999; Petford *et al.*, 2000) exhibiting
475 a density contrast with their host rocks.

476 In contrast to static surveys, dynamic gravimetric observations allow spatio-temporal mass
477 changes to be tracked. Dynamic gravimetric studies investigate how the subsurface architecture
478 changes over time and is usually performed by measuring variations in gravity across a network of
479 survey points (e.g., Fig. 7c) or, in a few exceptional cases, by installing a network of continuously
480 operating gravimeters. Dynamic observations demand one-to-two orders of magnitude higher data
481 precision (i.e. to a few μGal where $1 \mu\text{Gal} = 10^{-8} \text{ m/s}^2$) compared to static surveys, making them an
482 elaborate and time-consuming exercise. However, dynamic gravity data yields important insights
483 into the source processes behind non-tectonic volcano and crustal deformation, particularly if
484 combined with surface deformation data (e.g., InSAR and GNSS) as subsurface mass and volume
485 changes can be employed to characterise the density of the material behind the stress changes (Figs
486 7c-f and 8) (e.g., Battaglia & Segall, 2004; Jachens & Roberts, 1985; Poland & Carbone, 2016).
487 There are also cases where volcano unrest, due either to magma intrusion into a ductile host rock or
488 to volatile migration at shallow depths, does not result in resolvable surface deformation; in these
489 scenarios, gravity data have provided vital clues about subsurface processes otherwise hidden from
490 conventional monitoring techniques (e.g., Gottsmann *et al.*, 2006; Gottsmann *et al.*, 2007; Miller *et*
491 *al.*, 2017).

492 Whilst static and dynamic gravimetric observations offer considerable insight into the
493 structure and dynamics of magma plumbing systems, care must be exercised when collecting and
494 interpreting gravity data from active magmatic areas where seasonal variations in hydrothermal
495 systems, aquifers, or the vadose zone can influence subsurface mass distribution (e.g., Hemmings *et*

496 *al.*, 2016). These seasonal changes can, in some cases, result in data aliasing artefacts and inhibit
497 the quantification of deeper-seated magmatic processes (e.g., Gottsmann *et al.*, 2005; Gottsmann *et*
498 *al.*, 2007).

499

500 ***Observations***

501 Gravimetric investigations have been at the heart of studies into the subsurface structure of active
502 and ancient magma plumbing systems for more than 80 years (e.g., Carbone *et al.*, 2017; Van Camp
503 *et al.*, 2017). Using techniques initially designed for imaging salt domes, silicic plutons were the
504 first components of magma plumbing systems to be examined using gravimetry because their low
505 density relative to surrounding rocks produces clear, negative gravity anomalies of ~10 to ~40 mGal
506 amplitude (e.g., Reich, 1932; Bucher, 1944; Bott, 1953). Gravity data have been instrumental in the
507 investigation of upper-crustal, silicic magma plumbing systems, helping to reveal: (i) the 3D
508 geometry of plutons by allowing floor morphologies (e.g., flat-floored or wedge-shaped) to be
509 determined (e.g., Vigneresse *et al.*, 1999; Petford *et al.*, 2000); and (ii) how plutons are constructed,
510 for example, by the amalgamation of multiple intrusions fed from depth by dykes (e.g., Vigneresse,
511 1995). Furthermore, recent high-precision static surveys over active silicic volcanoes have enabled
512 detailed modelling of the sub-volcanic magma plumbing system, commonly demonstrating the
513 occurrence of vertically extensive, transcrustal magma bodies (Figs 7a and b) (e.g., Gottsmann *et*
514 *al.*, 2008; del Potro *et al.*, 2013; Saxby *et al.*, 2016; Miller *et al.*, 2017). In addition to examining
515 silicic magma plumbing systems, negative gravity anomalies with typical amplitudes of up to 60
516 mGal and up to 100 km wavelength can be associated with, and provide insight into, the geometry
517 and size of silicic ash-flow calderas (e.g., Eaton *et al.*, 1975; Masturyono *et al.*, 2001). Positive
518 gravity anomalies with amplitudes of up to 30 mGal and wavelengths of up to 20 km are
519 commonly identified at mafic volcanoes and likely result from dense intrusive complexes (e.g.,
520 Rymer & Brown, 1986).

Dynamic gravity observations have provided unprecedented insight into the evolution of magma plumbing systems over timescales of seconds to decades, including: (i) the characterisation of multi-year lava lake dynamics (e.g., Poland & Carbone, 2016); (ii) mass budgets of magma intrusions (e.g., Fig. 8) (e.g., Battaglia *et al.*, 1999; Jousset *et al.*, 2000; Rymer *et al.*, 2005; Bonforte *et al.*, 2007; Tizzani *et al.*, 2009); (iii) shallow hydrothermal fluid flow processes induced by deeper magmatic unrest (e.g., Battaglia *et al.*, 2006; Gottsmann *et al.*, 2007; Miller *et al.*, 2017); and (iv) parameters of magmatic geothermal reservoirs (e.g., Hunt & Bowyer, 2007; Sofyan *et al.*, 2011). For example, using data from a network of continuously recording gravimeters, Carbone *et al.*, (2013) calculated the density of the Kilauea lava lake as $950 \pm 300 \text{ kg m}^{-3}$, i.e. similar to and potentially less than that of water, suggesting that the magma column within the upper portions of the volcanic edifice is gas-rich. Because density and volatile content are critical controls on magma rheology, identification of a gas-rich magma column and lava lake at Kilauea is crucial to modelling and understanding convection and eruption dynamics (Carbone *et al.*, 2013).

Implications and integration

The advent of data-rich geodetic observations from satellite-remote sensing (e.g., InSAR), in conjunction with spatio-temporal gravity studies, provides unprecedented opportunities to characterise magma plumbing system dynamics and the driving mechanisms behind volcano deformation. At Long Valley caldera, for example, a residual gravity increase of more than $60 \text{ } \mu\text{Gal}$ between 1982 and 1999 indicates a mass addition at depth (Battaglia *et al.*, 1999). Joint inversion of InSAR and gravity data from Long Valley derives a best fit-source density of 2509 kg m^{-3} and is indicative of a magmatic intrusion (Fig. 8) (Tizzani *et al.*, 2009). At the deforming Laguna del Maule volcanic centre, Chile, multi-year InSAR and dynamic gravity records demonstrate that uplift and extension above an inflating sill-like reservoir at $\sim 5 \text{ km}$ depth promoted migration of hydrothermal fluids along a fault to shallow ($1\text{--}2 \text{ km}$) depths (Miller *et al.*, 2017). Alternatively, although no ground deformation is observed at Tenerife, Spain, deconvolution of dynamic gravity

547 into a shallow and deep gravity field provides evidence of unrest (Prutkin *et al.*, 2014). The gravity
548 data suggest hybrid processes have generated the unrest, whereby fluids were released and migrated
549 upward along deep-rooted faults from an intrusion at ~9 km beneath the summit of Teide Volcano
550 (Prutkin *et al.*, 2014). Overall, combining ground deformation and gravimetric observations has
551 highlighted complex processes both within magma reservoirs (e.g., mass addition by magma input,
552 density decrease by volatile exsolution, or density increase by crystallisation; Figs 7c-f) and in the
553 surrounding host rock (e.g., migration of magmatic fluids, phase changes in hydrothermal systems).
554 Key to a better understanding of the processes governing these magma plumbing system and
555 volcano deformation dynamics is the integration of gravimetric and geodetic data with other
556 geophysical data (e.g., seismicity or magnetotellurics) and petrological data. Coupled with
557 advanced numerical modelling, such multi-parameter studies promise exciting new insights into the
558 inner workings of sub-volcanic magma plumbing systems (e.g., Currenti *et al.*, 2007; Hickey *et al.*,
559 2016; Currenti *et al.*, 2017; Gottsmann *et al.*, 2017; Miller *et al.*, 2017).

560

561 **2.5. Resolving magma plumbing system structure with electromagnetic methods**

562 ***Techniques***

563 Electromagnetic (EM) methods probe subsurface electrical resistivity or its inverse, i.e. electrical
564 conductivity. Spatial variations in resistivity control the position, strength, and geometry of local
565 electrical eddy currents and the magnetic fields they produce. These electrical eddy currents are
566 induced by time-varying, naturally occurring magnetic fields external to Earth, which forms the
567 basis of the magnetotelluric (MT) technique, or by controlled sources. Monitoring these decaying
568 electrical and magnetic fields with passive MT techniques therefore allows the subsurface resistivity
569 distribution to be inferred. Controlled source methods generally probe only the shallow subsurface,
570 but MT has a greater depth range as it uses longer-period signals to penetrate deeper. The signals
571 propagate diffusively, which means EM methods typically have a lower resolution than seismic
572 techniques. However, melt, magma, and magmatic hydrothermal fluids are generally considerably

573 less resistive than solid rock and can thus easily be detected by EM methods, which are sensitive to
574 conductive materials (e.g., Whaler & Hautot, 2006; Wannamaker *et al.*, 2008; Desissa *et al.*, 2013;
575 Comeau *et al.*, 2015). EM methods, particularly MT, have therefore been used extensively to study
576 magmatic systems in various tectonic settings.

577 MT equipment, data acquisition, and processing is described by Simpson & Bahr (2005) and
578 Ferguson (2012). Measured field variations have very low amplitudes, meaning equipment needs to
579 be positioned and installed carefully to reduce vibrational (e.g., from wind, vegetation, or vehicles)
580 and electrical (e.g., from power lines) noise. If data are recorded synchronously at a second, less
581 noisy site, remote reference methods can be used to improve the data quality (e.g., Gamble *et al.*,
582 1979). One further problem is that small-scale resistivity anomalies in the shallow subsurface
583 generate galvanic (non-inductive) effects that distort MT data. The distortion is identified and
584 corrected for, which may involve using controlled source transient electromagnetic data to ensure
585 complete removal (e.g. Sternberg *et al.*, 1988), at the same time as assessing whether the data can
586 be modelled with a one-, two- or three-dimensional resistivity structure (e.g. Jones, 2012). Failure
587 to remove galvanic distortion can result in models having resistivity features at the wrong depth.
588 For example, there has been controversy as to whether a conductor beneath Vesuvius Volcano, Italy
589 is caused by a deep (~8–10 km depth) magma reservoir (Di Maio *et al.*, 1998) or a shallow brine
590 layer (Manzella *et al.*, 2004). All of these factors can be a significant problem when using MT to
591 study magmatic systems, especially on volcanic islands.

592 The relationship between MT data and subsurface resistivity is strongly non-linear meaning
593 that inversion is fundamentally non-unique and computationally expensive (e.g., Bailey, 1970;
594 Parker, 1980; Weaver, 1994). Most practical algorithms for inverting MT data obtain a unique
595 result by minimising a combination of misfit to the data and a measure of model roughness (e.g.,
596 Constable *et al.*, 1987). This approach poorly delimits how magma is distributed in the subsurface,
597 whether it is in sills, dykes, or larger reservoirs (Johnson *et al.*, 2016). Whilst MT data are sensitive
598 to the top surface of a conductor, its base may not be detected because conductive material reduces

599 the penetration depth of the signal. Sensitivity analysis is used to ascertain the model features
600 required to fit the MT data, which allows a conductor to be confined to a certain depth range and
601 thereby constrains its base (e.g., Desissa *et al.*, 2013). Furthermore, if the resistivity of a conductor
602 can be inferred, its conductance (i.e. a product of thickness and conductivity) can be used to
603 determine its thickness (e.g., Comeau *et al.*, 2016).

604

605 ***Observations***

606 EM induction surveys have been conducted on most major sub-aerial volcanoes and magmatic
607 systems; only a few will be mentioned here to illustrate the type information on magma plumbing
608 systems that has been obtained. MT data have been used to image several low resistivity features in
609 the central Andes, particularly beneath the uplifting (10–15 mm/yr) Volcán Uturuncu, Bolivia (Fig.
610 9a) (Comeau *et al.*, 2015; Comeau *et al.*, 2016). The deepest of these bodies has resistivities of <3
611 Ω m, has a top contact at ~15–20 km depth (i.e. it is shallowest beneath Uturuncu), likely has a
612 thickness of >6 km, and extends E-W for ~170 km (Fig. 9) (Comeau *et al.*, 2015; Comeau *et al.*,
613 2016). This large-scale structure is interpreted to be the Altiplano-Puna magma body (APMB),
614 which has been identified in other geophysical datasets (e.g., Fig. 7a) (e.g., gravimetry, del Potro *et*
615 *al.*, 2013), with its low resistivity attributed to the presence of at least 20% andesitic melt and/or
616 magma. Extending from the top of the APMB towards the surface are several vertical, narrow (<10
617 km wide), low resistivity ($<10 \Omega$ m) zones that coincide with areas of seismicity and negative
618 gravity anomalies (Fig. 9). These zones likely reflect a network of dykes and upper crustal magma
619 reservoirs (Jay *et al.*, 2012; del Potro *et al.*, 2013; Comeau *et al.*, 2015; Comeau *et al.*, 2016).

620 Monitoring of magmatic systems can also be undertaken by both time-lapse and continuous
621 EM measurement. For example, MT data collected immediately after the 1977–1978 eruption at
622 Usu volcano, Japan revealed a conductive zone ($<100 \Omega$ m) beneath the summit that probably
623 corresponded to intruded magma. By 2000, MT data revealed that this conductive body had become
624 resistive (500–1000 Ω m) as the intrusion cooled, from 800°C to 50°C, and crystallised

625 (Matsushima *et al.*, 2001). Continuous MT monitoring of Sakurajima volcano, Japan between May
626 2008 and July 2009 revealed temporal changes in resistivity of $\pm 20\%$, some of which correlated to
627 periods of surface deformation and were inferred to reflect mixing between groundwater and
628 volatiles exsolved from an underlying magma body (Aizawa *et al.*, 2011). Continuous MT
629 monitoring at La Fournaise, Réunion Island recorded apparent resistivity decreases associated with
630 the large 1998 eruption, which were attributed to the injection of a N-S striking dyke (Wawrzyniak
631 *et al.*, 2017).

632 Several EM studies have focussed on magma plumbing systems at divergent margins,
633 including mid-ocean ridges and continental rifts. For example, at the fast-spreading East Pacific
634 Rise, a ~10 km wide, sub-vertical conductor, slightly displaced from the ridge axis and connected to
635 a deep, broad conductive zone was interpreted as a channel efficiently transporting melt to the base
636 of the crust (Baba *et al.*, 2006; Key *et al.*, 2013). Imaging of a crustal conductor for the first time
637 beneath a slow-spreading ridge, i.e. the Reykjanes ridge in the Atlantic Ocean, suggests that magma
638 injection into crustal reservoirs is intermittent but rapid (MacGregor *et al.*, 1998; Heinson *et al.*,
639 2000). Conversely, slow-spreading continental rifting in the Dabbahu magma segment, Afar,
640 Ethiopia appears to be underlain by a large conductor, either at the top of the mantle or straddling
641 the Moho, containing more melt ($>300 \text{ km}^3$) than is intruded into the magma plumbing system
642 during a typical rifting episode (Desissa *et al.*, 2013). The volume of this large conductor implies it
643 is a long-lived feature that could source magmatic activity for tens of thousands of years (Desissa *et*
644 *al.*, 2013).

645

646 ***Implications and integration***

647 It is clear from MT studies of the APMB that other geophysical techniques aid and/or corroborate
648 data interpretation (Fig. 9) (e.g., Comeau *et al.*, 2015; Comeau *et al.*, 2016). Over the last two
649 decades, numerous geophysical studies have been applied to examine magma and melt distribution
650 beneath various portions of the East African Rift, providing an excellent opportunity to test how

651 different techniques and data can be integrated. For example, extensive zones of melt beneath the
652 Afar region in Ethiopia inferred from MT data by Desissa *et al.*, (2013) is supported by: (i) the
653 occurrence of coincident, low P-wave velocity (down to 7.2 km s^{-1}) zones identified using from
654 analysis of seismic Pn waves that propagate along the Moho (Stork *et al.*, 2013); (ii) surface wave
655 studies that reveal lower crustal areas in magmatic domains with low S-wave velocities ($\sim 3.2 \text{ km s}^{-1}$) (Guidarelli *et al.*, 2011); and (iii) high anisotropic V_p/V_s ratios and low amplitude receiver
656 functions, which are indicative of the presence of melt (Hammond *et al.*, 2011; Hammond, 2014).
657 Similarly, crustal conductors along the northern flanks of the Main Ethiopian Rift, interpreted to
658 represent melt/magma (Whaler & Hautot, 2006; Samrock *et al.*, 2015; Hübner *et al.*, 2018), coincide
659 with locations where receiver functions either have amplitudes too low to interpret or indicate high
660 V_p/V_s values (Dugda *et al.*, 2005; Stuart *et al.*, 2006). Electrical anisotropy can be inferred directly
661 from MT data consistent with a two-dimensional subsurface resistivity distribution (Padilha *et al.*,
662 2006; Hamilton *et al.*, 2006). Large amounts of electrical anisotropy were found in the lower crust
663 beneath Quaternary magmatic segments in Afar, Ethiopia, where there is also significant crustal
664 seismic anisotropy (see Fig. 11 of Ebinger *et al.*, 2017); oriented melt-filled pockets are the
665 probable cause of both.

667 Although EM methods can image subsurface conductors that are interpreted to represent
668 magma bodies or zones of partial melt (i.e. crystal mushes), additional information is required to
669 determine their composition, volume, and/or melt fraction. However, there are several challenges in
670 inverting measured bulk resistivities to recover this information. Two-phase mixing laws predict
671 bulk resistivity is primarily a function of melt resistivity and geometry in the rock matrix when the
672 fluid phase has low resistivity, as in the case of partial melt. Well-connected melt gives a lower bulk
673 resistivity than isolated melt pockets, for the same melt fraction and resistivity (e.g. Hashin &
674 Shtrikman, 1963; Roberts & Tyburezy, 1999; Schmeling, 1986). Whilst resistivities of basaltic and
675 rhyolitic melts have been measured in laboratory experiments (e.g., Laumonier *et al.*, 2015; Guo *et al.*, 2016), they are strongly dependent on temperature, pressure, silica, sodium and water content,

677 making extrapolation uncertain. The web-based SIGMELTS tool can, however, be used to predict
678 melt and bulk resistivities for a wide range of compositions and conditions (Pommier & Le Trong,
679 2011). Importantly, petrological and geochemical characterisation of eruptive products can help
680 inform interpretations of associated, subsurface conductors but it is difficult to ascertain either
681 whether their composition reflects the current magma/melt present in the plumbing system or
682 whether melt pockets are interconnected. These large uncertainties in melt resistivity and the
683 requirement to make assumptions about its geometry make direct inference of melt fraction
684 difficult. Nonetheless, information from laboratory studies, petrology, and geochemistry aids
685 interpreting resistivity anomalies in magmatic regions (see review by Pommier, 2014).

686

687 **2.6. Imaging ancient magma plumbing systems in seismic reflection data**

688 ***Techniques***

689 Over the last two decades, major advances have been made in imaging deep crustal melt beneath
690 active volcanic terrains using P- and S-wave tomographic data (e.g., Yellowstone, Husen *et al.*,
691 2004; Mt. St. Helens, Lees, 2007; Hawaii, Okubo *et al.*, 1997). These data image deep (>7 km),
692 often laterally extensive (up to 20 km), sill-like magma reservoirs (e.g., Paulatto *et al.*, 2012).
693 However, like many geophysical and geodetic techniques applied to study active magma plumbing
694 systems, these data typically lack the spatial resolution to resolve the detailed geometry of pathways
695 transporting magma to the Earth's surface. Active source seismic reflection data, which have a
696 spatial resolution of metres-to-decametres down to depths of ~5 km, can provide unprecedented
697 images of and insights into the geometry and dynamics of shallow-level, crystallised, magma
698 plumbing systems (e.g., Fig. 10) (e.g., Planke *et al.*, 2000; Smallwood & Maresh, 2002; Thomson &
699 Hutton, 2004; Cartwright & Hansen, 2006; Jackson *et al.*, 2013; Magee *et al.*, 2016; Schofield *et*
700 *al.*, 2017). Whilst seismic reflection data are traditionally used to find and assist in the production of
701 hydrocarbons in sedimentary basins (Cartwright & Huuse, 2005), we here discuss and support its
702 application to volcanological problems.

703 Acquiring active source seismic reflection data involves firing acoustic energy (i.e. seismic
704 waves) into the subsurface and measuring the surface arrival times (i.e. the travel-time) of reflected
705 energy. Processing of these arrival time data allows reconstruction of the location and geometry of
706 the geological interfaces from which acoustic energy was reflected. Mafic intrusive igneous rocks
707 are generally well-imaged in seismic reflection data because they typically have greater densities
708 ($>2.5 \text{ g/cm}^3$) and acoustic velocities (i.e. $>4000 \text{ m/s}$) than encasing sedimentary strata; these
709 differences result in a high acoustic impedance contrast, causing more seismic energy to be
710 reflected back to the surface compared to low acoustic impedance boundaries (Smallwood &
711 Maresh, 2002; Brown, 2004). In contrast, silicic igneous rocks have similar acoustic properties to
712 encasing sedimentary strata, meaning that felsic intrusions are rarely imaged in seismic reflection
713 data (Mark *et al.*, 2017; Rabbel *et al.*, 2018). Furthermore, because reflection seismology relies on
714 the return of acoustic energy to the surface, seismic reflection data favourably image mafic, sub-
715 horizontal-to-moderately inclined intrusions (e.g., sills, inclined sheets, and laccoliths; Smallwood
716 & Maresh, 2002; Jackson *et al.*, 2013; Magee *et al.*, 2016). Sub-vertical dykes reflect only a limited
717 amount of acoustic energy back to the surface and are thus typically poorly imaged in seismic
718 reflection data (e.g., Smallwood & Maresh, 2002; Planke *et al.*, 2005; Thomson, 2007; Wall *et al.*,
719 2010; Eide *et al.*, 2017a; Phillips *et al.*, 2017).

720

721 ***Observations***

722 Sills and inclined sheets are commonly observed in seismic reflection data as laterally
723 discontinuous, high-amplitude reflections, which may cross-cut the host rock strata (Fig. 10) (e.g.,
724 Symonds *et al.*, 1998, Smallwood & Maresh, 2002; Planke *et al.*, 2005, Magee *et al.*, 2015). Many
725 of the sills and inclined sheets imaged in seismic reflection data are, however, expressed as tuned
726 reflection packages, whereby discrete reflections from the top and base contacts interfere on their
727 return to the surface and cannot be distinguished (e.g., Figs 10 and 11a) (e.g., Smallwood &
728 Maresh, 2002; Peron-Pinvidic *et al.*, 2010; Magee *et al.*, 2015; Eide *et al.*, 2017a; Rabbel *et al.*,

2018). It is therefore difficult to assess either intrusion thicknesses, or to detect whether imaged sills are composite bodies made of numerous, stacked, thin sheets. Either way, subtle vertical offsets and corresponding amplitude variations of sill reflections can often be mapped, defining linear structures that radiate out from either the central, deepest portions of sills or areas where underlying intrusions intersect the sill (e.g., Schofield *et al.*, 2012a; Magee *et al.*, 2014; Magee *et al.*, 2016). These structures are interpreted to relate to magma flow indicators such as intrusive steps, broken bridges, and magma fingers (e.g., Schofield *et al.*, 2010; Schofield *et al.*, 2012b; Magee *et al.*, 2018).

A recurring observation from seismic reflection-based studies of extinct and buried intrusive systems is that complexes of interconnected sills and inclined sheets, which may cover $>3 \times 10^6$ km², can dominate magma plumbing systems (e.g., Fig. 10b) (e.g., Svensen *et al.*, 2012, Magee *et al.*, 2016). Importantly, where buried volcanic edifices are imaged in seismic reflection data, they rarely appear to be underlain by ‘magma chambers’ (i.e. a spheroidal or ellipsoidal body of now-crystallised magma). Instead, these imaged volcanoes commonly appear laterally offset from genetically related sills and/or laccoliths that are inferred to represent their feeder reservoirs (e.g., Fig. 10b) (Magee *et al.*, 2013a; McLean *et al.*, 2017). The geometry, location, and connectivity of these intrusions, which can represent magma storage sites and conduits to the surface, are often heavily influenced by both the host rock structure and lithology (see review by Magee *et al.*, 2016). For example, magma may flow along pronounced discontinuities (e.g., bedding) or within specific stratigraphic units (e.g., coal) for considerable distances, occasionally climbing to higher stratigraphic levels by instigating deformation of the host rock or by exploiting pre-existing faults (e.g., Jackson *et al.*, 2013; Magee *et al.*, 2016; Schofield *et al.*, 2017; Eide *et al.*, 2017b). It is clear from seismic reflection data that shallow-level tabular intrusions are commonly accommodated by roof uplift to form a flat-topped or dome-shaped forced fold (e.g., Figs 11a and b) (e.g., Trude *et al.*, 2003; Hansen & Cartwright, 2006; Jackson *et al.*, 2013; Magee *et al.*, 2013b). Moreover, if the age of reflections onlapping onto these intrusion-induced forced folds can be ascertained, the timing and

755 to some extent the duration of magmatic activity can be determined (e.g., Trude *et al.*, 2003;
756 Hansen & Cartwright, 2006; Magee *et al.*, 2014; Reeves *et al.*, 2018). Although most seismic-based
757 studies examine intrusions within sedimentary basins, saucer-shaped sills and laterally extensive
758 sill-complexes emplaced into crystalline basement rock are also imaged (e.g., Ivanic *et al.*, 2013;
759 McBride *et al.*, 2018). Lastly, seismic reflection data can also be used to image the internal structure
760 of layered ultramafic-mafic intrusions (e.g., the Bushveld Layered Intrusion, Malehmir *et al.*, 2012)
761 and, in some instances, identify dykes (e.g., Fig. 11c) (e.g., Wall *et al.*, 2010; Abdelmalak *et al.*,
762 2015; Bosworth *et al.*, 2015; Phillips *et al.*, 2017).

763

764 ***Implications and integration***

765 Despite being limited in terms of their spatial resolution (typically a few tens of metres) and ability
766 to image steeply dipping features (i.e. dykes), they provide unprecedented snapshots into the final
767 3D structure of magma plumbing systems. Beyond quantifying the structure and connectivity of
768 magma plumbing systems, seismic-based studies have shown that: (i) magma flow patterns mapped
769 across entire sill-complexes indicate they can transport melt from source to surface over great
770 lateral (>100's km) and vertical distances (10's km), potentially without significant input from
771 dykes (Fig. 10a) (e.g., Thomson & Hutton, 2004; Cartwright & Hansen, 2006; Magee *et al.*, 2014;
772 Magee *et al.*, 2016; Schofield *et al.*, 2017); and (ii) a variety of elastic and inelastic mechanisms can
773 accommodate host rock deformation during magma emplacement, meaning that the location and
774 size of ground deformation does not necessarily equal that of the forcing intrusion (e.g., Jackson *et al.*,
775 2013, Magee *et al.*, 2013b). Importantly, observations from seismic reflection data highlight that
776 the lateral dimension should be considered when modelling the transit of magma in the crust, posing
777 problems for the widely held and simple assumption that magma simply travels vertically from melt
778 source to eruption site.

779 Seismic-based studies have also shown that direct comparison to active deformation
780 structures can be informative. For example, through comparing mapped lava flows and structures

781 associated with the Alu dome to similar features observed in seismic reflection data (see section
782 2.6), Magee *et al.*, (2017) concluded that the shallow-level sill likely has a saucer-shaped, as
783 opposed to the sill-like tabular morphology inferred from an episode of deformation measured using
784 InSAR (Figs 3c and d). Despite its benefits, it is important to remember that seismic reflection data
785 typically reveal only the final geometry of the magma plumbing system. There thus remains a
786 challenge in using these data to understand areas where deformation captures potentially transient,
787 active processes, rather than structures resulting from (multiple) periods of intrusion and cooling
788 (Reeves *et al.*, 2018). One potential and exciting way forward is the development of Virtual
789 Reflection Seismic Profiling, by which microseismicity at active volcanoes may be used to image
790 magma reservoirs and subsurface structure in 4D (Kim *et al.*, 2017). Although challenges exist in
791 dataset integration, the imaging power afforded by modern seismic reflection data thus presents a
792 unique opportunity to further unite field-, petrological-, geochemical-, and other geophysical-based
793 analyses within more realistic structural frameworks (e.g., Figs 3, 11a and b). In our view, however,
794 seismic reflection data are under-utilized in igneous research, remaining an unfamiliar technique to
795 many Earth Scientists in the volcanic and magmatic community.

796

797 **2.7. Rock magnetism**

798 ***Technique***

799 Whilst seismic reflection data provide unique 3D images of ancient magma plumbing systems,
800 which can be used to infer magma flow patterns across entire intrusion networks, we commonly
801 lack sufficient data (e.g., boreholes) to test seismic-based hypotheses. It is therefore critical to
802 compare seismic interpretations to field analogues where magma flow patterns, emplacement
803 mechanics, and intrusion evolution can be investigated via other techniques. In this section, we
804 examine how rock magnetic analyses can be used to systematically study magnetic mineralogy and
805 petrofabrics, thereby illuminating the structure and history of igneous intrusions.

806 There are two principal types of rock magnetic study; magnetic remanence and magnetic
807 susceptibility, where the total magnetisation (M) of a rock is the sum of the magnetic remanence
808 (M_{rem}) and the induced magnetisation (M_{ind}), which is a product of the susceptibility (K) and
809 applied field strength (H) (Dunlop & Özdemir, 2001). Remanence carries a geological record of the
810 various magnetisations acquired over time and is central to palaeomagnetic studies. However, we
811 focus on magnetic fabric analysis, which relies on measurements of the anisotropy of magnetic
812 susceptibility (AMS). The AMS signal of a rock carries information from all constituent grains.
813 Although mineral phases that have a paramagnetic behaviour (i.e. they are weakly attracted to
814 externally applied magnetic fields) volumetrically dominate most igneous rocks (e.g., olivine,
815 clinopyroxene, biotite), ferromagnetic mineral phases (e.g., titanomagnetite) are highly susceptible
816 to magnetization and therefore tend to dominate K (e.g., Dunlop & Özdemir, 2001; Biedermann *et*
817 *al.*, 2014). Magnetic fabrics therefore typically reflect the preferential orientation of
818 crystallographic axes (i.e. crystalline anisotropy), the shape-preferred orientation of individual
819 crystals (i.e. shape anisotropy), and/or the alignment of closely spaced crystals (i.e. distribution
820 anisotropy) belonging to Fe-bearing silicate and oxide phases (e.g., Voight & Kinoshita, 1907;
821 Graham, 1954; Hrouda, 1982; Tarling & Hrouda, 1993; Dunlop & Özdemir, 2001). The principal
822 axes of the magnetic fabrics measured by AMS can thus be related to the orientation, shape, and
823 distribution of individual grains (i.e. the petrofabric) (e.g., Fig. 12a).

824 Regardless of whether mineral phases crystallise early or late, whereby their orientation and
825 distribution typically mimics the earlier silicate framework, it is expected that the initial petrofabric
826 developed in intrusive rocks will likely be sensitive to alignment of crystals during primary magma
827 flow. However, it is also critical to recognise that later magmatic processes (e.g., convection and
828 melt extraction) and syn- or post-emplacement tectonic deformation can modify or overprint
829 primary magma flow fabrics during intrusion, solidification (i.e. mush development), or sub-solidus
830 conditions (e.g., Borradaile & Henry, 1997; Bouchez, 1997; O'Driscoll *et al.*, 2015; Kavanagh *et*
831 *al.*, 2018). Whilst anisotropy of magnetic susceptibility (AMS) can thus rapidly and accurately

832 detect weak or subtle mineral alignments within igneous intrusions, which may be attributable to
833 magmatic and/or tectonic processes, evaluating the origin and evolution of petrofabric development
834 requires additional information (e.g., Borradaile & Henry, 1997; Bouchez, 1997). For example,
835 shape-preferred orientation analyses and comparison to visible flow indicators (e.g., intrusive steps
836 and bridge structures) allow magma flow axes and directions that have been inferred from magnetic
837 fabrics to be verified (e.g., Launeau & Cruden, 1998; Callot *et al.*, 2001; Magee *et al.*, 2012a). For a
838 useful précis of AMS-related magnetic theory in igneous rocks, the reader is referred to early works
839 by Balsey & Buddington (1960) and Khan (1962), and more recent summaries provided by Martín-
840 Hernández *et al.*, (2004), O'Driscoll *et al.*, (2008), and O'Driscoll *et al.*, (2015).

841 The principle behind AMS relies on the measurement of the bulk susceptibility (K_m) of a
842 single sample in different orientations to determine the susceptibility anisotropy tensor, which
843 relates the induced magnetisation (M_{ind}) to the applied field (H) in three dimensions (Tarling &
844 Hrouda, 1993). The orientation and magnitude of the eigenvectors and eigenvalues of this tensor
845 define an ellipsoid with three principal axes; the long axis of the ellipsoid, K_1 , defines the magnetic
846 lineation and the short axis, K_3 , defines the normal (i.e. the pole) to the magnetic foliation plane
847 (K_1 – K_2 ; Fig. 12a) (Stacy *et al.*, 1960; Khan, 1962; Tarling & Hrouda, 1993). In order to interpret
848 magnetic fabrics, it is important to determine the mineralogy of the phases carrying the magnetic
849 signal because the composition, grainsize, and distribution of magnetically dominant minerals (e.g.,
850 titanomagnetite) can control fabric orientation (e.g., Hargreaves *et al.*, 1991; Stephenson, 1994;
851 Dunlop & Özdemir, 2001). In addition to primary crystallographic and textural controls on
852 magnetic fabrics, subsequent oxidation of remaining melt and secondary hydrothermal alteration
853 can affect the magnetic mineralogy and, thereby, the AMS signal (e.g., Trindade *et al.*, 2001;
854 Stevenson *et al.*, 2007a). A variety of rock magnetic experiments are thus required to determine the
855 magnetic mineralogy. The most widely used method involves measuring susceptibility, and thereby
856 behaviour of magnetic materials, at varying temperatures ranging from -200°C to 700°C (i.e.
857 thermomagnetic analysis *sensu* Orlický, 1990; Hrouda *et al.*, 1997). For example, paramagnetic

858 materials (e.g., biotite) follow the Curie-Weiss law, whereby their susceptibility drops
859 hyperbolically with increasing temperature. In contrast, the thermomagnetic curve of ferromagnetic
860 materials (e.g., titanomagnetite) displays little change in susceptibility with temperature, apart from
861 when characteristic crystallographic transitions occur (e.g., the Curie point for pure magnetite at
862 ~580°C, Petrovský & Kapička, 2006) temperature. To determine the grainsize of ferromagnetic
863 fraction in the magnetic susceptibility signal, the hysteretic property of the magnetisation is
864 important (Dunlop, 2002). Other rock magnetic experiments (e.g., anisotropy of anhysteretic
865 remanent magnetism (AARM) can be conducted to further isolate the relative importance of
866 different paramagnetic and ferromagnetic phases (e.g., McCabe *et al.*, 1985; Richter & van der
867 Pluijm, 1994; Kelso *et al.*, 2002).

868

869 ***Observations***

870 Having established the magnetic mineralogy, AMS fabrics can be interpreted. Even in weakly
871 anisotropic igneous rocks (i.e. visually isotropic), particularly sheet intrusions, it is now accepted
872 that the magnetic lineation and foliation can provide information on magma migration (e.g., flow
873 direction) or regional and local strain (e.g., Hrouda, 1982; Knight & Walker, 1988; Rochette *et al.*,
874 1992; Bouchez, 1997; Tauxe *et al.*, 1998; Callot *et al.*, 2001; Féménias *et al.*, 2004; Magee *et al.*,
875 2012a). For example, comparisons to other indicators of magma flow (e.g., intrusive steps and
876 visible mineral alignments) in sheet intrusions have shown that magnetic lineations commonly
877 parallel the magma flow (e.g., Knight & Walker, 1988; Cruden & Launeau, 1994; Callot *et al.*,
878 2001; Magee *et al.*, 2012a), whilst imbrication of elongate crystals induced by simple shear at
879 intrusion margins define the sense of magma flow (Fig. 12b) (e.g., Knight & Walker, 1988;
880 Hargraves *et al.*, 1991; Stephenson, 1994; Geoffroy *et al.*, 2002; Féménias *et al.*, 2004).
881 Alternatively, contact-parallel magnetic fabrics generated during the formation and inflation of
882 magma lobes can be used to determine flow and emplacement dynamics, even if other evidence for
883 the presence of magma lobes is lacking (e.g., Fig. 12c) (Cruden *et al.*, 1999; Stevenson *et al.*,

2007a; Magee *et al.*, 2012b). Identifying changes in fabric orientation within or between individual sheet intrusions is also important because these variations suggest that deformation, imparted by either the emplacement of adjacent magma bodies or tectonic processes, did not significantly modify magma emplacement fabrics (e.g., Clemente *et al.*, 2007).

Post solidification textural modification and the possibility of overlap in tectonic and magmatic strain fields during protracted emplacement is a particular complication when studying granitoid and gabbroic plutons (e.g., Mamtani *et al.*, 2013; O'Driscoll *et al.*, 2015; Cheadle *et al.*, 2017). In fact, most early studies of granitoid emplacement using AMS, in conjunction with many other structural analysis tools, concluded that tectonic strain was the main source of subtle fabrics (e.g., Brun *et al.*, 1990; Bouchez, 1997; de Saint-Blanquat & Tikoff 1997; Neves *et al.*, 2003; Mamtani *et al.*, 2005). Although primary magma flow fabrics in granitic and gabbroic plutons may thus be overprinted, the magnetic fabrics characterised by AMS can still provide fundamental insights into emplacement mechanics (e.g., Stevenson *et al.*, 2007a; Petronis *et al.*, 2012) and magma/mush evolution (e.g., formation of layering; O'Driscoll *et al.*, 2015).

Implications and integration

Overall, AMS has provided vital magma flow and evolution information that has helped to understand mafic and silicic magma plumbing systems (e.g., Knight & Walker, 1988; Ernst & Baragar, 1992; Glen *et al.*, 1997; Aubourg *et al.*, 2008; Petronis *et al.*, 2013; Petronis *et al.*, 2015). Critical insights emanating from these AMS studies have revealed that: (i) flow trajectories predicted by classic emplacement models (e.g., for ring dykes and cone sheets) are not always consistent with measured AMS fabrics and supporting data, which thereby call into question the application of such models (e.g., Stevenson *et al.*, 2007b; Magee *et al.*, 2012a); (ii) lateral magma flow is recorded in many shallow, planar intrusions associated with volcanic magma plumbing systems (e.g., Ernst & Baragar, 1992; Cruden & Laneau, 1994; Cruden *et al.*, 1999; Herrero-Bervera *et al.*, 2001; Magee *et al.*, 2012a; Petronis *et al.*, 2013; Petronis *et al.*, 2015); and (iii)

910 plutons, particularly those with a granitic composition, commonly consist of incrementally
911 emplaced magma pulses that often develop lobate geometries (e.g., Fig. 12c) (e.g., Stevenson et al.,
912 2007a). Analysing AMS fabrics from layered mafic-ultramafic intrusions can also provide evidence
913 for magma reservoir processes, including crystal settling, or post-cumulus modification of crystal
914 mushes (O'Driscoll *et al.*, 2008; O'Driscoll *et al.*, 2015). Importantly, AMS and related analyses
915 provide robust, testable, and repeatable methods to constrain subtle shape and crystallographic
916 orientations of crystals in igneous rocks. Rock magnetic instrumentation technology continues to
917 advance with better automation of measurement protocols, sensitivity of measurements, and a
918 greater ability to unravel contributors to the AMS signal. The direction and scope of these
919 developments are improving the holistic integration of AMS with other structural, microstructural,
920 geophysical, petrological and geochemical techniques, promising to advance our understanding of
921 magmatism and crustal evolution.

922

923 **3. Future advances**

924 Our understanding of magma plumbing system structure and evolution has been significantly
925 enhanced by the geophysical techniques described above. We have demonstrated that there is scope
926 for advancement within individual methodologies and through the integration of different
927 techniques, particularly involving the synthesis of geophysical, petrological, and geochemical data.
928 In this section, we discuss two new techniques that will potentially revolutionize our understanding
929 of magma plumbing systems. We also briefly discuss how integration of geophysical data with
930 numerical modelling can enhance our knowledge of reservoir construction and evolution.

931

932 **3.1. Full-Waveform Inversion**

933 ***Technique***

934 We have demonstrated that seismic reflection data can provide unique insight into the 3D structure
935 of magma plumbing systems (e.g., see review by Magee *et al.*, 2016). In addition to using seismic

936 reflection data to image the subsurface, we can also invert the measured travel-times of reflected
937 acoustic energy to model subsurface P-wave velocities. Full-waveform inversion (FWI) is a rapidly
938 developing technology using active source seismic data to generate models that reproduce both the
939 travel-times and full waveform of the arriving wavefield, thereby matching observed seismic data
940 (Tarantola, 1984). Because FWI considers the full wavefield, as opposed to conventional techniques
941 that only model travel-times, it is a technique capable of recovering high-resolution models of
942 subsurface P-wave velocities and other physical properties (Warner *et al.*, 2013; Routh *et al.*, 2017).
943 The FWI technique begins with a best-guess starting velocity model for the subsurface geology,
944 which is then iteratively updated using a local linearized inversion until the observed seismic data is
945 matched (Virieux & Operto, 2009). FWI is much more computationally expensive than travel-time
946 tomography, as a full-physics implementation of the wave equation is required to generate the
947 predicted seismic data at all energy source and receiver locations for each iteration (Routh *et al.*,
948 2017). FWI, however, has the advantage of being able to resolve much finer-scale structure than
949 conventional techniques.

950

951 ***Observations***

952 To date, 3D FWI has principally been applied within the petroleum sector to obtain high-resolution
953 velocity models that can be used to improve depth-migrated (i.e. travel-time is converted to depth in
954 metres) reflection images of petroleum reservoirs and their overburden (Sirgue *et al.*, 2010; Vigh *et al.*,
955 2010; Warner *et al.*, 2013; Kapoor *et al.*, 2013; Routh *et al.*, 2017). FWI can also produce
956 interpretable, quantitative models of physical properties of rocks in the subsurface that can be
957 related directly to compaction, permeability, and overpressure as measured in subsurface boreholes
958 (Lazaratos *et al.*, 2011; Mancini *et al.*, 2015). Of relevance here is that mafic intrusions, which
959 appear as high-amplitude reflections in seismic reflection data (Figs 10 and 11a), are recovered as
960 high-velocity features in FWI velocity models (Fig. 13) (Mancini *et al.*, 2015; Kalincheva *et al.*,
961 2017). For example, successful application of 3D FWI to a marine ocean bottom seismometer

962 dataset acquired across the Endeavour segment of the Juan de Fuca Ridge led to generation of a
963 velocity model that had a resolution up to four times greater than travel-time tomography (Morgan
964 *et al.*, 2016). Within this new, high-resolution velocity model, several velocity anomalies were
965 identified and interpreted to indicate localized magma recharge of the axial reservoir, induced
966 seismogenic cracking, and increased permeability (Arnoux *et al.*, 2017).

967

968 ***Implications and integration***

969 Active magma plumbing systems comprise a complex network of interconnected conduits and
970 reservoirs with variable geometries and sizes, which likely contain magmatic vapour-rich, liquid-
971 rich, and mush-zones (Christopher *et al.*, 2015). These intrusions will all be associated with reduced
972 P-wave velocities, which could be resolved in high-resolution, 3D FWI datasets as supported by
973 successes in the fine-scale imaging of: (i) low-velocity gas clouds (Warner *et al.*, 2013); (ii) axial
974 reservoirs at an oceanic spreading centre (Arnoux *et al.*, 2017); (iii) relatively narrow, low-velocity
975 fault zones within an antiform (Morgan *et al.*, 2013); and (iv) a subduction zone using 2D FWI
976 (Kamei *et al.*, 2012). A suite of synthetic tests has been performed to investigate whether 3D FWI
977 could be applied to better understand magma plumbing systems (Morgan *et al.*, 2013). These tests
978 indicate that it is possible to recover high-resolution models of P-wave velocity beneath volcanoes,
979 which can then be used to better determine where magma/mush is stored beneath the surface. In
980 particular, these synthetic tests suggest that FWI could be used to: (i) distinguish between
981 continuous zones of mush and individual magma reservoirs; (ii) image sills and conduits of magma
982 and/or fluids that are a few 10s metres across (Fig. 13); and (iii) image the deeper (lower-crustal)
983 part of the magma system. We therefore consider that 3D FWI affords an unprecedented
984 opportunity to obtain high-resolution images of actual magma plumbing systems beneath active
985 volcanoes. To this end, the ongoing PROTEUS (Plumbing Reservoirs Of The Earth Under
986 Santorini) experiment was specifically designed to use 3D FWI to investigate the Santorini magma
987 plumbing system (Hooft *et al.*, 2017).

988

989 **3.2. Unmanned Aerial Vehicle photogrammetry**

990 *Technique*

991 Despite major advances in satellite-based remote sensing systems and aeromagnetic surveys, very
992 high-resolution (i.e., mm–cm scale ground sampling distance) imagery of dykes and other igneous
993 intrusions has been limited to low altitude aerial photography. This in turn has created a critical
994 scale gap in intrusion studies, which range from <1 mm at thin section scale to the metres to 100's
995 of metres scale provided by outcrop analysis, conventional remote sensing, and geophysical data.
996 Fortunately, the emerging capability of unmanned aerial vehicle (UAV) photogrammetry fills this
997 gap (e.g., Eisenbeiss, 2009; Westoby *et al.*, 2012; Bemis *et al.*, 2015; Eide *et al.*, 2017b). It is also
998 noteworthy that several studies have demonstrated that digital photogrammetry can deliver high
999 quality datasets with accuracies similar to more established laser scanning techniques (e.g., Leberl
1000 *et al.*, 2010; Hodgetts, 2013; Thiele *et al.*, 2015).

1001 The basic setup required to carry out UAV (or drone) photogrammetry is commercially
1002 available and relatively inexpensive, comprising a fixed wing or rotary wing UAV, a digital camera,
1003 and access to a suitable digital photogrammetry software package (e.g., Agisoft Photoscan Pro,
1004 Pix4Dmapper Pro, VisualSFM). UAV photogrammetry combines a simple and cost-effective
1005 method to acquire geospatially referenced, overlapping digital aerial images, from which structure-
1006 from-motion algorithms can generate spatial 3D datasets (Bemis *et al.*, 2014; Vollgger & Cruden,
1007 2016). Such an approach can be used for high spatial resolution mapping of all types of well-
1008 exposed igneous intrusions. The resulting data greatly enhance the effectiveness of traditional field
1009 mapping, particularly the characterisation of contact relationships and internal and external structure
1010 (e.g., fractures, fabrics, and phase distributions) of intrusive rocks, complementing AMS and
1011 petrological analyses.

1012

1013 *Observations*

1014 Aa photogrammetric workflow was applied to examine a swarm of 5 cm to 1 m wide Palaeogene
1015 dolerite and dacite dykes exposed on coastal outcrops at Bingie Bingie Point, SE Australia (Fig.
1016 14). The orthophotograph of the entire wave-cut platform shows the distribution of the Palaeogene
1017 dolerite and dacite dykes and their Devonian host rock lithologies, including a prominent
1018 moderately NE-dipping aplite dyke (Fig. 14a). Linear ENE-WSW linear terrain features pick out
1019 the traces of dyke-parallel joints (Fig. 14a). The Palaeogene dykes trend 063° parallel to a major set
1020 of joints in the country rock that likely formed contemporaneously with syn-dyking extension (Fig.
1021 14b). Subsidiary joint sets trend NNW-SSE, sub-perpendicular to the Palaeogene dykes, N-S and E-
1022 W (Fig. 14b). The Palaeogene dykes display considerable structural complexity such as bridge
1023 structures, intrusive steps and apophyses (Fig. 14c). Where present, the steps mostly occur where
1024 dykes cross country rock contacts (e.g., the aplite-tonalite contact in the NE; Fig. 14c).

1025

1026 ***Implications and integration***

1027 Data such as the orthophotograph collected at Bingie Bingie Point indicate that high-resolution
1028 structural and lithological mapping and measurement can be carried out much more rapidly than by
1029 traditional survey methods (e.g., plane table or grid mapping). However, the use of conventional
1030 RGB cameras restricts the resulting image data to reflected visible light. Future applications will
1031 include the deployment of multispectral and hyperspectral sensors (infrared to short wave infrared
1032 to thermal infrared) as well as potential field geophysical or geodetic instruments (e.g., Sparks,
1033 2012). A further challenge for UAV applications in many countries concerns the regulatory
1034 framework around the use of drones for research. The global trend is moving to require non-
1035 recreational UAV operators to have remotely piloted aircraft licences and for the associated
1036 organisation to be certified for UAV operations. Innovations in sensor types and design, attachment
1037 of geophysical instruments, machine learning, and integration with complementary techniques such
1038 as AMS will open up new avenues for UAV applications in the study of magma plumbing systems.

1039

1040 **3.4. Numerical modelling of magma reservoir processes constrained by geophysical data**

1041 Geophysical imaging of both active and ancient magma plumbing systems is delivering new
1042 insights into the 3D geometry of reservoirs, the timing and rates of melt and magma transport, the
1043 pathways followed by magmas as they ascend through the crust, and typical stored melt fractions in
1044 mushes. These data can be used to constrain and calibrate numerical models of reservoir processes.
1045 Numerical models are used ubiquitously to understand and predict the behaviour of other
1046 subsurface crustal reservoirs, such as hydrocarbon reservoirs, groundwater resources, and targets for
1047 geological CO₂ storage (e.g., Chen *et al.*, 2003; Class *et al.*, 2009; Dean & Chen, 2011). However,
1048 there has been relatively little focus to date on developing numerical models for magma/mush
1049 reservoirs. Yet such models can integrate across different data sources and types, provide
1050 quantitative estimates of rates, volumes and timescales, and provide a framework for data
1051 interpretation. For example, numerical modelling of heat transfer within the plumbing system at
1052 Okmok Volcano in Alaska, which was informed by analytical models of geodetic data and
1053 estimated magma compositions of erupted material, allowed estimation of the role magma injection,
1054 crystallisation, and degassing processes had on volume changes over time (Caricchi *et al.*, 2014).
1055 Numerical thermal modelling has also helped interpret seismic data from the Soufrière Hills
1056 Volcano, Montserrat, suggesting higher melt fraction in the underlying magma reservoir than was
1057 inferred from seismic data alone (Paulatto *et al.*, 2012). More recent numerical models focus on
1058 crystal mushes, evaluating melt transport and reaction at low melt fractions, and these show that
1059 temperature and melt fraction in mushes can be decoupled; i.e. maximum temperature occurs close
1060 to the centre of the reservoir but maximum melt fraction occurs close to the top (Solano *et al.*,
1061 2014). This decoupling impacts how seismic velocities and electrical conductivities will be
1062 modified within the mush (Solano *et al.*, 2014). Other numerical models show the important role
1063 played by exsolution, crystallisation, and the viscoelastic response of the crust in driving magma
1064 mobilisation in and eruption from shallow reservoirs (e.g., Degruyter & Huber, 2014; Parmigiani *et*
1065 *al.*, 2016), as well as providing insights into the mixing mechanisms of melt and crystals in mushes

1066 (Bergantz *et al.*, 2015). However, most models to date have a lower dimensionality (zero dimension
1067 box models, or one/two dimensions) and capture only a small subset of the key physical and
1068 chemical processes that are likely to occur in crustal magma reservoirs or crystal mushes.
1069 Moreover, few studies have integrated modelling with geophysical data (cf. Gutierrez *et al.*, 2013).
1070 This is in marked contrast to the 3D modelling routinely undertaken of other crustal reservoirs (e.g.,
1071 hydrocarbon reservoirs), which is commonly integrated with and delimited by geophysical data.
1072 There is thus significant scope for improved, and integrated, numerical modelling of crustal magma
1073 reservoirs.

1074

1075 **4. Conclusions**

1076 Determining the structure of magma plumbing systems is critical to understanding where melt and
1077 magma is stored in the crust, which can influence the location of volcanic eruptions and economic
1078 ore deposits, providing an important framework for interpreting the physical and chemical evolution
1079 of magma from petrological and geochemical datasets. Geophysical techniques have revealed
1080 unique insights into the architecture of active and ancient magma plumbing systems, which when
1081 integrated with traditional structural, petrological and geochemical results has yielded exciting
1082 advances in our understanding of magmatic processes. However, divisions between communities
1083 applying these methodologies still exist, contributing to diverging views on the nature of magma
1084 plumbing systems. To help promote collaboration, we have reviewed a range of geophysical
1085 techniques and discussed how they could be integrated with structural, petrological and
1086 geochemical datasets to answer outstanding questions in the volcanological community. In
1087 particular, we demonstrate how a range geophysical techniques can be applied to track melt
1088 migration in near real-time, map entire intrusion networks in 3D, examine magma emplacement
1089 mechanics, and understand the evolution of crystal mushes. For example, Interferometric Synthetic
1090 Aperture Radar (InSAR) allows measurement of the development of active magmatic systems by
1091 successive intrusion, the vertical and lateral movements of magma, and the relationship between

1092 magma plumbing system dynamics and eruption. Seismicity beneath volcanoes can, when the
1093 magma interacts dynamically with the host rock, illuminate in high-resolution the time and spatial
1094 scales of the motion of magma and hydrothermal fluids. Seismic imaging of magma plumbing
1095 systems allows the spatial distribution of melt and magma to be determined whilst the inclusion of
1096 anisotropy within seismic techniques even allows sub-seismic wavelength features to be identified.
1097 Gravimetry can characterise the distribution and redistribution of mass (e.g., magma) in the
1098 subsurface over high spatial and temporal resolutions, helping to reveal the structure and
1099 composition of magma plumbing systems and the source(s) of volcano deformation.
1100 Electromagnetic methods, particularly magnetotellurics, can identify fluids within magmatic
1101 systems (e.g., melt, magma, and hydrothermal fluids). Seismic reflection data provide
1102 unprecedented 3D images of ancient magma plumbing systems and has revealed that laterally
1103 extensive, interconnected networks of sills and inclined sheets can play a pivotal role in transporting
1104 magma through the crust to eruption sites potentially located >100 km away from the melt source.
1105 Rock magnetics can provide fabric data pertaining to magma flow, deformation or crystallisation.
1106 All these methodologies discussed have provided unique insights into the structure of igneous
1107 intrusions and, through integration with petrological and geochemical datasets, are beginning to
1108 help unravel the entire evolution of magma plumbing systems. In addition to the ongoing
1109 application and advancement of these geophysical techniques, emerging methodologies look set to
1110 radically improve our understanding of magma plumbing systems. For example, full-waveform
1111 inversion can image and characterise physical properties across plumbing systems at an
1112 unprecedented resolution, whereas unmanned aerial vehicle photogrammetry provides a tool for
1113 high spatial resolution of outcrop scale intrusions that bridges the scale gap between seismic
1114 reflection data and traditional mapping of magma plumbing systems. The geophysical techniques
1115 discussed also provide critical constraints on input parameters for numerical modelling. Overall, we
1116 consider that the future of magma plumbing system studies will benefit greatly from the synthesis
1117 of geophysics and more traditional petrological and geochemical approaches.

1118

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1126 **6. References**

- 1127 Abdelmalak, M. M., Andersen, T. B., Planke, S., Faleide, J. I., Corfu, F., Tegner, C., Shephard, G.
1128 E., Zastrozhnov, D., Myklebust, R. (2015). The ocean-continent transition in the mid-Norwegian
1129 margin: Insight from seismic data and an onshore Caledonian field analogue. *Geology* **43**, 1011-
1130 1014.
- 1131 Aizawa, K., Kandam, W., Ogawa, Y., Iguchi, M., Yokoo, A., Yakiwara, H., Sugano, T. (2011).
1132 Temporal changes in electrical resistivity at Sakurajima volcano from continuous magnetotelluric
1133 observations. *Journal of Volcanology and Geothermal Research* **199**, 165-175.
- 1134 Annen, C. (2011). Implications of incremental emplacement of magma bodies for magma
1135 differentiation, thermal aureole dimensions and plutonism–volcanism relationships. *Tectonophysics*
1136 **500**, 3-10.
- 1137 Annen, C., Blundy, J. D., Sparks, R. S. J. (2006). The genesis of intermediate and silicic magmas in
1138 deep crustal hot zones. *Journal of Petrology* **47**, 505-539.
- 1139 Annen, C., Blundy, J. D., Leuthold, J., Sparks, R. S. J. (2015). Construction and evolution of
1140 igneous bodies: Towards an integrated perspective of crustal magmatism. *Lithos* **230**, 206-221.
- 1141 Arnoux G. M., Toomey, D., Hooft, E., Wilcock, W., Morgan, J., Warner, M., VanderBeek, B.
1142 (2017). Seismic evidence that black smoker heat flux is rate-limited by crustal permeability.
1143 *Geophysical Research Letters* **44**, 1687-1695.

1144 Aubourg, C., Tshoso, G., le Gall, B., Bertrand, H., Tiercelin, J. -J., Kampunzu, A. B., Dymment, J.,
 1145 Modisi, M. (2008). Magma flow revealed by magnetic fabric in the Okavango giant dyke swarm,
 1146 Karoo igneous province, northern Botswana. *Journal of Volcanology and Geothermal Research*
 1147 **170**, 247-261.

1148 Baba, K., Chave, A. D., Evans, R. L., Hirth, G., Mackie, R. L. (2006). Mantle dynamics beneath the
 1149 East Pacific Rise at 17°S: Insights from the Mantle Electromagnetic and Tomography (MELT)
 1150 experiment. *Journal of Geophysical Research* **111**, B02101.

1151 Bailey, R. C. (1970). Inversion of the geomagnetic induction problem. *Proceedings of the Royal*
 1152 *Society* **315**, 185-194.

1153 Bagnardi, M., Amelung, F., Poland, M. P. (2013). A new model for the growth of basaltic shields
 1154 based on deformation of Fernandina volcano, Galápagos Islands. *Earth and Planetary Science*
 1155 *Letters* **377**, 358-366.

1156 Balsley, J. R., & Buddington, A. F. (1960). Magnetic susceptibility anisotropy and fabric of some
 1157 Adirondack granites and orthogneisses. *American Journal of Science* **A258**, 6-20.

1158 Bannister, S., Reyners, M., Stuart, G. & Savage, M. (2007). Imaging the Hikurangi subduction
 1159 zone, New Zealand, using teleseismic receiver functions: crustal fluids above the forearc mantle
 1160 wedge. *Geophysical Journal International* **169**, 602-616.

1161 Battaglia, M., Roberts, C. & Segall, P. (1999). Magma intrusion beneath Long Valley caldera
 1162 confirmed by temporal changes in gravity. *Science* **285**, 2119-2122.

1163 Battaglia, M. & Segall, P. (2004). The Interpretation of Gravity Changes and Crustal Deformation
 1164 in Active Volcanic Areas. *Pure and Applied Geophysics* **161**, 1453-1467.

1165 Battaglia, M., Troise, C., Obrizzo, F., Pingue, F. & De Natale, G. (2006). Evidence for fluid
 1166 migration as the source of deformation at Campi Flegrei caldera (Italy). *Geophysical Research*
 1167 *Letters* **33**, L01307.

1168 Bemis, S. P., Micklethwaite, S., Turner, D., James, M. R., Akciz, S., Thiele, S., Bangash, H. A.
 1169 (2014). Ground-based and UAV-Based photogrammetry: A multi-scale, high-resolution mapping
 1170 tool for structural geology and paleoseismology. *Journal of Structural Geology* **69**, 163-178.
 1171 Bergantz, G. W., Schleicher, J. M., Burgisser, A. (2015) Open-system dynamics and mixing in
 1172 magma mushes. *Nature Geoscience* **8**, 793-796.
 1173 Berryman, J. G. (1980). Long-wavelength propagation in composite elastic media II. Ellipsoidal
 1174 inclusions. *The Journal of the Acoustical Society of America* **68**, 1820-1831.
 1175 Bersi, M., Saibi, H. & Chabou, M. C. (2016). Aerogravity and remote sensing observations of an
 1176 iron deposit in Gara Djebilet, southwestern Algeria. *Journal of African Earth Sciences* **116**, 134-
 1177 150.
 1178 Biedermann, A. R., Pettke, T., Reusser, E., Hirt, A. M. (2014). Anisotropy of magnetic
 1179 susceptibility in natural olivine single crystals. *Geochemistry, Geophysics, Geosystems* **15**, 3051–
 1180 3065.
 1181 Biggs, J., Anthony, E. Y. & Ebinger, C. J. (2009). Multiple inflation and deflation events at Kenyan
 1182 volcanoes, East African Rift. *Geology* **37**, 979-982.
 1183 Biggs, J., Bastow, I. D., Keir, D. & Lewi, E. (2011). Pulses of deformation reveal frequently
 1184 recurring shallow magmatic activity beneath the Main Ethiopian Rift. *Geochemistry, Geophysics,*
 1185 *Geosystems* **12**, Q0AB10.
 1186 Biggs, J., Ebmeier, S. K., Aspinall, W. P., Lu, Z., Pritchard, M. E., Sparks, R. S. J., Mather, T. A.
 1187 (2014). Global link between deformation and volcanic eruption quantified by satellite
 1188 imagery. *Nature Communications* **5**, 3471.
 1189 Biggs, J. & Pritchard, M. E. (2017). Global volcano monitoring: what does it mean when volcanoes
 1190 deform?. *Elements* **13**, 17-22.
 1191 Bonforte, A., Carbone, D., Greco, F. & Palano, M. (2007). Intrusive mechanism of the 2002 NE-rift
 1192 eruption at Mt Etna (Italy) modelled using GPS and gravity data. *Geophysical Journal International*
 1193 **169**, 339-347.

1194 Borradaile, G. J. & Henry, B. (1997). Tectonic applications of the magnetic susceptibility and its
 1195 anisotropy. *Earth Science Reviews* **42**, 49-93.

1196 Bosworth, W., Stockli, D. F. & Helgeson, D. E. (2015). Integrated outcrop, 3D seismic, and
 1197 geochronologic interpretation of Red Sea dike-related deformation in the Western Desert, Egypt–
 1198 The role of the 23Ma Cairo “mini-plume”. *Journal of African Earth Sciences* **109**, 107-119.

1199 Bott, M. H. P. (1953). Negative gravity anomalies over acid "intrusions" and their relation to the
 1200 structure of the Earth's crust. *Geological Magazine* **90**, 257-267.

1201 Bouchez, J. L. (1997). Granite is never isotropic: An introduction to AMS studies of granitic rocks.
 1202 In: Bouchez, J. L., Hutton, D. H. W., Stephens, W. E. (eds) Granite: From Segregation of Melt to
 1203 Emplacement Fabrics 8: pp. 95-112. Dordrecht: Kluwer Academic.

1204 Brown, A. R. (2004). *Interpretation of three-dimensional seismic data*. Oklahoma, USA: AAPG
 1205 and SEG.

1206 Brun, J. P., Gapais, D., Cogne, J. P., Ledru, P. & Vigneresse, J. L. (1990). The Flamanville granite
 1207 (northwest France): an unequivocal example of a syntectonically expanding pluton. *Geological*
 1208 *Journal* **25**, 271-286.

1209 Bucher, W. H. (1944). Discussion in Romberg, F., and Barnes, V. E. Correlation of gravity
 1210 observations with the geology of the Smoothingiron granite mass, Llano County, Texas. *Geophysics*
 1211 **9**, 79-93.

1212 Callot, J. -P., Geoffroy, L., Aubourg, C., Pozzi, J., Mege, D. (2001). Magma flow directions of
 1213 shallow dykes from the East Greenland volcanic margin inferred from magnetic fabric studies.
 1214 *Tectonophysics* **335**, 313-329.

1215 Carbone, D., Poland, M. P., Patrick, M. R. & Orr, T. R. (2013). Continuous gravity measurements
 1216 reveal a low-density lava lake at Kīlauea Volcano, Hawai‘i. *Earth and Planetary Science Letters*
 1217 **376**, 178-185.

1218 Carbone, D., Poland, M. P., Diament, M. & Greco, F. (2017). The added value of time-variable
1219 microgravimetry to the understanding of how volcanoes work. *Earth-Science Reviews* **169**, 146-
1220 179.

1221 Caricchi, L., Biggs, J., Annen, C., Ebmeier, S. (2014) The influence of cooling, crystallisation and
1222 re-melting on the interpretation of geodetic signals in volcanic systems. *Earth and Planetary
1223 Science Letters* **388**, 166-174.

1224 Cartwright, J. & Hansen, D. M. (2006). Magma transport through the crust via interconnected sill
1225 complexes. *Geology* **34**, 929-932.

1226 Cartwright, J. & Huuse, M. (2005). 3D seismic technology: the geological ‘Hubble’. *Basin
1227 Research* **17**, 1-20.

1228 Cashman, K. V. & Sparks, R. S. J. (2013). How volcanoes work: A 25 year perspective. *Geological
1229 Society of America Bulletin* **125**, 664-690.

1230 Cashman, K. V., Sparks, R. S. J. & Blundy, J. D. (2017). Vertically extensive and unstable
1231 magmatic systems: a unified view of igneous processes. *Science* **355**, p.eaag3055.

1232 Chaussard, E. & Amelung, F. (2013) Depth of magma storage in volcanic arcs: testing the influence
1233 of regional parameters using a global data compilation. In EGU General Assembly Conference
1234 Abstracts (Vol. 15).

1235 Cheadle, M. J., & Gee, J. S. (2017). Quantitative textural insights into the formation of gabbro in
1236 mafic intrusions. *Elements* **13**, 409-414.

1237 Chen, Y., Durlofsky, L. J., Gerritsen, M., Wen, X. H. (2003). A coupled local–global upscaling
1238 approach for simulating flow in highly heterogeneous formations. *Advances in Water Resources* **26**,
1239 1041-1060.

1240 Chmielowski, J., Zandt, G. & Haberland, C. (1999). The central Andean Altiplano-Puna magma
1241 body. *Geophysical Research Letters* **26**, 783-786.

1242 Chouet, B., Dawson, P., Martini, M. (2008). Shallow-conduit dynamics at Stromboli Volcano, Italy,
1243 imaged from waveform inversions. In: Lane, S. J., Gilbert, J. S. (Eds.), *Fluid Motions in Volcanic*

1244 Conduits: A Source of Seismic and Acoustic Signals. *Geological Society of London Special*
1245 *Publication* **307**, 57-84.

1246 Christopher, T. E., Blundy, J. D., Cashman, K., Cole, P., Edmonds, M., Smith, P. J., Sparks, R. S. J.
1247 & Stinton, A. (2015). Crustal-scale degassing due to magma system destabilization and magma-gas
1248 decoupling at Soufrière Hills Volcano, Montserrat. *Geochemistry Geophysics Geosystems* **16**, 2797-
1249 2811.

1250 Chouet, B. & Matoza, R. S. (2013). A Multi-decadal view of seismic methods for detecting
1251 precursors of magma movement and eruption. *Journal of Volcanology and Geothermal Research*
1252 **252**, 108-175.

1253 Class, H., Ebigbo, A., Helmig, R., Dahle, H. K., Nordbotten, J. M., Celia, M. A., Audigane, P.,
1254 Darcis, M., Ennis-King, J., Fan, Y., Flemisch, B. (2009). A benchmark study on problems related to
1255 CO₂ storage in geologic formations. *Computational Geosciences* **13**, 409-434.

1256 Clemente, C. S., Amorós, E. B. & Crespo, M. G. (2007). Dike intrusion under shear stress: effects
1257 on magnetic and vesicle fabrics in dikes from rift zones of Tenerife (Canary Islands). *Journal of*
1258 *Structural Geology* **29**, 1931-1942.

1259 Comeau, M. J., Unsworth, M. J., Ticona, F., Sunagua, M. (2015). Magnetotelluric images of
1260 magma distribution beneath Volcán Uturuncu, Bolivia: Implications for magma dynamics. *Geology*
1261 **43**, 243-246.

1262 Comeau, M. J., Unsworth, M. J., Cordell, D. (2016). New constraints on the magma distribution
1263 and composition beneath Volcán Uturuncu and the southern Bolivian Altiplano from
1264 magnetotelluric data. *Geosphere* **12**, 1391-1421.

1265 Constable, S. C., Parker, R. L., Constable, C. G. (1987). Occam's inversion: A practical algorithm
1266 for generating smooth models from electromagnetic sounding data. *Geophysics* **52**, 289-300.

1267 Cruden, A. R. & Launeau, P. (1994). Structure, magnetic fabric and emplacement of the Archean
1268 Lebel Stock, SW Abitibi greenstone belt. *Journal of Structural Geology* **16**, 677-691.

1269 Cruden, A. R., Tobisch, O. T., Launeau, P. (1999). Magnetic fabric evidence for conduit fed
1270 emplacement of a tabular granite: Dinkey Creek Pluton, central Sierra Nevada Batholith, California.
1271 *Journal of Geophysical Research* **104**, 10,511-10,531.

1272 Cruden, A. R. & Weinberg, R. F. (2018). Mechanisms of magma transport and storage in the lower
1273 and middle crust – magma segregation, ascent and emplacement. In: Burkhardt, S. (Ed). *Volcanic
1274 and Igneous Plumbing Systems*, Elsevier, Amsterdam, In Press.

1275 Currenti, G., Del Negro, C. & Ganci, G. (2007). Modelling of ground deformation and gravity
1276 fields using finite element method: an application to Etna volcano. *Geophysical Journal
1277 International* **169**, 775-786.

1278 Currenti, G., Napoli, R., Coco, A. & Privitera, E. (2017). Effects of hydrothermal unrest on stress
1279 and deformation: insights from numerical modeling and application to Vulcano Island (Italy).
1280 *Bulletin of Volcanology* **79**, 28.

1281 de Saint Blanquat, M. & Tikoff, B. (1997). Development of magmatic to solid-state fabrics during
1282 syntectonic emplacement of the Mono Creek Granite, Sierra Nevada Batholith. In *Granite: from
1283 segregation of melt to emplacement fabrics*, Springer Netherlands, 231-252.

1284 Dean, O. S. & Chen, Y. (2011). Recent progress on reservoir history matching: a review.
1285 *Computational Geosciences* **15**, 152-221.

1286 DeGruyter, W. & Huber, C. (2014). A model for eruption frequency of upper crustal silicic magma
1287 chambers. *Earth and Planetary Science Letters* **403**, 117-130.

1288 del Potro, R., Díez, M., Blundy, J., Gottsmann, J. & Camacho, A. (2013). Diapiric ascent of silicic
1289 magma beneath the Bolivian Altiplano. *Geophysical Research Letters* **40**, 2044-2048.

1290 Delph, J. R., Ward, K. M., Zandt, G., Ducea, M. N., Beck, S. L. (2017). Imaging a magma
1291 plumbing system from MASH zone to magma reservoir. *Earth and Planetary Science Letters* **457**,
1292 313-324.

1293 Desissa, M., Johnson, N. E., Whaler, K. A., Hautot, S., Fisseha, S., Dawes, G. (2013). A mantle
1294 magma reservoir beneath an incipient mid-ocean ridge in Afar, Ethiopia. *Nature Geoscience* **6**, 861-

1295 865.

1296 Di Maio, R., Mauriello, P., Patella, D., Petrillo, Z., Piscitelli, S., Siniscalchi, A. (1998). Electric and
1297 electromagnetic outline of the Mount Somma–Vesuvius structural setting. *Journal of Volcanology*
1298 *and Geothermal Research* **82**, 219-238.

1299 Dixon, T. H. (1991). An introduction to the Global Positioning System and some geological
1300 applications. *Reviews of Geophysics* **29**, 249-276.

1301 Drew, J., White, R. S., Tilmann, F., Tarasewicz, J. (2013) Coalescence microseismic mapping.
1302 *Geophysical Journal International* **195**, 1773-1785.

1303 Dugda, M. T., Nyblade, A. A., Julia, J., Langston, C. A., Ammon, C. J., Simiyu, S. (2005). Crustal
1304 structure in Ethiopia and Kenya from receiver function analysis: Implications for rift development
1305 in eastern Africa. *Journal of Geophysical Research* **110**, B01303.

1306 Dunlop, D. J. (2002). Theory and application of the Day plot (Mrs/Ms versus Hcr/Hc) 1.
1307 Theoretical curves and tests using titanomagnetite data. *Journal of Geophysical Research: Solid*
1308 *Earth* **107**(B3).

1309 Dunlop, D. J. & Özdemir, Ö. (2001). Rock magnetism: fundamentals and frontiers (Vol. 3).
1310 Cambridge university press.

1311 Dvorak, J. J. & Dzurisin, D. (1997). Volcano geodesy: The search for magma reservoirs and the
1312 formation of eruptive vents. *Reviews of Geophysics* **35**, 343-384.

1313 Dzurisin, D. (2006). Volcano deformation: new geodetic monitoring techniques. Springer Science
1314 & Business Media.

1315 Eaton, G. P., Christiansen, R. L., Iyer, H. M., Pitt, A. D., Mabey, D. R., Blank, H. R., Zietz, I. &
1316 Gettings, M. E. (1975). Magma Beneath Yellowstone National park. *Science* **188**, 787.

1317 Ebinger, C., Keir, D., Ayele, A., Calais, E., Wright, T. J., Belachew, M., Hammond, J. O. S.,
1318 Campbell, E., Buck, W. R. (2008). Capturing magma intrusion and faulting processes during
1319 continental rupture: seismicity of the Dabbahu (Afar) rift. *Geophysical Journal International* **174**,
1320 1138-1152.

1321 Ebinger, C. J., Keir, D., Bastow, I. D., Whaler, K., Hammond, J. O., Ayele, A., Miller, M. S.,
 1322 Tiberi, C., Hautot, S. (2017). Crustal structure of active deformation zones in Africa: Implications
 1323 for global crustal processes. *Tectonics*.
 1324 Ebmeier, S. K., Elliott, J. R., Nocquet, J. M., Biggs, J., Mothes, P., Jarrín, P., Yépez, M., Aguaiza,
 1325 S., Lundgren, P., Samsonov, S. V. (2016). Shallow earthquake inhibits unrest near Chiles–Cerro
 1326 Negro volcanoes, Ecuador–Colombian border. *Earth and Planetary Science Letters* **450**, 283-291.
 1327 Ebmeier, S. K., Andrews, B. J., Araya, M. C., Arnold, D. W. D., Biggs, J., Cooper, C. & 10 others.
 1328 (2018). Synthesis of global satellite observations of magmatic and volcanic deformation:
 1329 implications for volcano monitoring & the lateral extent of magmatic domains. *Journal of Applied*
 1330 *Volcanology*, **7**, 2.
 1331 Eide, C. H., Schofield, N., Lecomte, I., Buckley, S. J. and Howell, J. A. (2017a). Seismic
 1332 interpretation of sill complexes in sedimentary basins: implications for the sub-sill imaging
 1333 problem. *Journal of the Geological Society of London*, jgs2017-096.
 1334 Eide, C. H., Schofield, N., Jerram, D. A., Howell, J. A. (2017b). Basin-scale architecture of deeply
 1335 emplaced sill complexes: Jameson Land, East Greenland. *Journal of the Geological Society of*
 1336 *London* **174**, 23-40.
 1337 Eisenbeiss, H. (2009). UAV Photogrammetry. PhD thesis. ETH Zurich.
 1338 Ernst, R. E. & Baragar, W. R. A. (1992). Evidence from magnetic fabric for the flow pattern of
 1339 magma in the Mackenzie giant radiating dyke swarm. *Nature* **356**, 511.
 1340 Féménias, O., Diot, H., Berza, T., Gauffriau, A., Demaiffe, D. (2004). Asymmetrical to symmetrical
 1341 magnetic fabric of dikes: Paleo-flow orientations and Paleo-stresses recorded on feeder-bodies from
 1342 the Motru Dike Swarm (Romania). *Journal of Structural Geology* **26**, 1401-1418.
 1343 Ferguson, I. J. (2012). Instrumentation and field procedures, in *The Magnetotelluric Method:*
 1344 *Theory and Practice*, ed. Chave, A. D. & Jones, A. G., Cambridge University Press.

1345 Field, L., Blundy, J. D., Brooker, R. A., Wright, T., Yirgu, G. (2012) Magma storage conditions
1346 beneath Dabbahu volcano (Ethiopia) constrained by petrology, seismicity and satellite geodesy.
1347 *Bulletin of Volcanology* **74**, 981-1004.

1348 Fournier, T. J., Pritchard, M. E., Riddick, S. N. (2010). Duration, magnitude, and frequency of
1349 subaerial volcano deformation events: New results from Latin America using InSAR and a global
1350 synthesis. *Geochemistry, Geophysics, Geosystems* **11**, Q01003.

1351 Gaillard, F. (2004). Laboratory measurements of electrical conductivity of hydrous and dry silicic
1352 melts under pressure. *Earth and Planetary Science Letters* **218**, 215-228.

1353 Galland, O. (2012). Experimental modelling of ground deformation associated with shallow magma
1354 intrusions. *Earth and Planetary Science Letters* **317**, 145-156.

1355 Gamble, T. D., Goubau, W. M., Clarke, J. (1979). Magnetotellurics with a remote magnetic
1356 reference. *Geophysics* **44**, 53-68.

1357 Garapic, G., Faul, U. H., Brisson, E. (2013). High-resolution imaging of the melt distribution in
1358 partially molten upper mantle rocks: evidence for wetted two-grain boundaries. *Geochemistry*
1359 *Geophysics Geosystems* **14**, 556-566.

1360 Geoffroy, L., Callot, J. P., Aubourg, C. & Moreira, M. (2002). Magnetic and plagioclase linear fabric
1361 discrepancy in dykes: a new way to define the flow vector using magnetic foliation. *Terra Nova* **14**,
1362 183-190.

1363 Gerst, A. & Savage, M. K. (2004). Seismic anisotropy beneath Ruapehu volcano: a possible eruption
1364 forecasting tool. *Science* **306**, 1543-1547.

1365 Glazner, A. F., Bartley, J. M. & Coleman, D. S. (2016). We need a new definition for "magma". *Eos*
1366 **97**.

1367 Glen, J. M., Renne, P. R., Milner, S. C., Coe, R. S. (1997). Magma flow inferred from anisotropy of
1368 magnetic susceptibility in the coastal Parana-Etendeka igneous province: Evidence for rifting before
1369 flood volcanism. *Geology* **25**, 1131-1134.

1370 Goes, S., Armitage, J., Harmon, N., Smith, H., Huisman, R. (2012). Low seismic velocities below

1371 mid-ocean ridges: Attenuation versus melt retention. *Journal of Geophysical Research* **117**, B12403.

1372 Gottsmann, J., Rymer, H. & Wooller, L. K. (2005). On the interpretation of gravity variations in the
1373 presence of active hydrothermal systems: Insights from the Nisyros Caldera, Greece. *Geophysical*
1374 *Research Letters* **32**, L23310.

1375 Gottsmann, J., Camacho, A. G., Tiampo, K. F. & Fernández, J. (2006). Spatiotemporal variations in
1376 vertical gravity gradients at the Campi Flegrei caldera (Italy): a case for source multiplicity during
1377 unrest?. *Geophysical Journal International* **167**, 1089-1096.

1378 Gottsmann, J., Carniel, R., Coppo, N., Wooller, L., Hautmann, S. & Rymer, H. (2007). Oscillations
1379 in hydrothermal systems as a source of periodic unrest at caldera volcanoes: Multiparameter insights
1380 from Nisyros, Greece. *Geophysical Research Letters* **34**, L07307.

1381 Gottsmann, J., Camacho, A. G., Marti, J., Wooller, L., Fernández, J., Garcia, A. & Rymer, H. (2008).
1382 Shallow structure beneath the Central Volcanic Complex of Tenerife from new gravity data:
1383 Implications for its evolution and recent reactivation. *Physics of the Earth and Planetary Interior* **68**
1384 212-230.

1385 Gottsmann, J., Blundy, J. D., Henderson, S., Pritchard, M. E. & Sparks, R. S. J. (2017).
1386 Thermomechanical Modeling of the Altiplano-Puna Deformation Anomaly: Multiparameter Insights
1387 into Magma Mush Reorganization. *Geosphere* **13**, 1042-1065.

1388 Graham, J. W. (1954). Magnetic anisotropy, an unexploited petrofabric element. *Geological Society*
1389 *of America Bulletin* **65**, 1257-1258.

1390 Gudmundsson, O., Brandsdottir, B., Menke, W. & Sigvaldason, G. E. (1994). The crustal magma
1391 chamber of the Katla volcano in south Iceland revealed by 2-D seismic undershooting. *Geophysical*
1392 *Journal International* **119**, 277-296.

1393 Guidarelli, M., Stuart, G., Hammond, J. O., Kendall, J. M., Ayele, A., Belachew, M. (2011).
1394 Surface wave tomography across Afar, Ethiopia: Crustal structure at a rift triple-junction zone.
1395 *Geophysical Research Letters* **38**, L24313.

1396 Guo, X., Zhang, L., Behrens, H., Ni, H. (2016). Probing the status of felsic magma reservoirs:
 1397 Constraints from the P – T – H_2O dependences of electrical conductivity of rhyolitic melt. *Earth and*
 1398 *Planetary Science Letters* **433**, 54-62.

1399 Gutierrez, F., Payacan, I., Gelman, S. E., Bachmann, O., Parada, M. A. (2013) Late-stage magma
 1400 flow in a shallow felsic reservoir: Merging the anisotropy of magnetic susceptibility record with
 1401 numerical simulations in La Gloria Pluton, central Chile. *Journal of Geophysical Research-Solid*
 1402 *Earth* **118**, 1984-1998.

1403 Hamilton, M. P., Jones, A. G., Evans, R. L., Evans, S., Fourie, C. J. S., Garcia, X., Mountford, A,
 1404 Spratt, J. E., SAMTEX MT Team. (2006). Electrical anisotropy of South African lithosphere
 1405 compared with seismic anisotropy from shear-wave splitting analyses. *Physics of the Earth and*
 1406 *Planetary Interiors* **158**, 226-239.

1407 Hamling, I. J., Wright, T. J., Calais, E., Bennati, L., Lewi, E. (2010). Stress transfer between
 1408 thirteen successive dyke intrusions in Ethiopia. *Nature Geoscience* **3**(10), 713-717.

1409 Hamling, I. J., Hreinsdottir, S. & Fournier, N. (2015). The ups and downs of the TVZ: Geodetic
 1410 observations of deformation around the Taupo Volcanic Zone, New Zealand. *Journal of*
 1411 *Geophysical Research: Solid Earth* **120**, 4667-4679.

1412 Hammer, S. (1945). Estimating ore masses in gravity prospecting. *Geophysics* **10**, 50-62.

1413 Hammond, W. & Humphreys, E. (2000a). Upper mantle seismic wave velocity: Effects of realistic
 1414 partial melt geometries. *Journal of Geophysical Research* **105**, 10975-10986.

1415 Hammond, W. & Humphreys, E. (2000b). Upper mantle seismic wave attenuation: Effects of
 1416 realistic partial melt distribution, *Journal of Geophysical Research* **105**, 10987-10999.

1417 Hammond, J. O. S., Kendall, J., Stuart, G. W., Keir, D., Ebinger, C. J., Ayele, A., Belachew, M.
 1418 (2011). The nature of the crust beneath the Afar triple junction: Evidence from receiver functions.
 1419 *Geochemistry Geophysics Geosystems* **12**, Q12004.

1420 Hammond, J. O. S. (2014). Constraining melt storage geometries beneath the Afar Depression,
 1421 Ethiopia from teleseismic receiver functions: The anisotropic H- κ stacking technique. *Geochemistry*

1422 *Geophysics Geosystems* **15**, 1316-1332.

1423 Hammond, J. O. S. & Kendall, J. (2016). Constraints on melt distribution from seismology: a case
 1424 study in Ethiopia. In: Wright, T. J., Ayele, A., Ferguson, D. J., Kidane, T., Vye-Brown, C. (eds.)
 1425 *Magmatic Rifting and Active Volcanism*. Geological Society, Special Publication, London **420**, 127-
 1426 147.

1427 Hansen, D. M. & Cartwright, J. (2006). The three-dimensional geometry and growth of forced folds
 1428 above saucer-shaped igneous sills. *Journal of Structural Geology* **28**, 1520-1535.

1429 Hargraves, R. B., Chan, C. Y., Johnson, D. (1991). Distribution anisotropy: the cause of AMS in
 1430 igneous rocks? *Geophysical Research Letters* **18**, 2193-2196.

1431 Hargraves, R. B., Briden, J. C., Daniels, B. A. (1999). Palaeomagnetism and magnetic fabric in the
 1432 Freetown Complex, Sierra Leone. *Geophysical Journal International* **136**, 705-713.

1433 Harmon, N., & Rychert, C. A. (2015). Seismic imaging of deep crustal melt sills beneath Costa Rica
 1434 suggests a method for the formation of the Archean continental crust. *Earth and Planetary Science*
 1435 *Letters* **430**, 140-148.

1436 Hashin, Z. & Shtrikman, S. (1962). A variational approach to the theory of the effective magnetic
 1437 permeability of multiphase materials. *Journal of Applied Physics* **33**, 3125-3131.

1438 Hawkesworth, C. J., Blake, S., Evans, P., Hughes, R., Macdonald, R., Thomas, L. E., Turner, S. P.,
 1439 Zellmer, G. (2000). Time scales of crystal fractionation in magma chambers—integrating physical,
 1440 isotopic and geochemical perspectives. *Journal of Petrology* **41**, 991-1006.

1441 Heinson, G., Constable, S., White, A. (2000). Episodic melt transport at mid-ocean ridges inferred
 1442 from magnetotelluric sounding. *Geophysical Research Letters* **27**, 2317-2320.

1443 Hemmings, B., Coco, A., Gottsmann, J., Whitaker, F. (2016). Investigating hydrological
 1444 contributions to volcano monitoring signals. *Geophysical Journal International* **207**, 259-273.

1445 Herrero-Bervera, E., Walker, G., Canon-Tapia, E., Garcia, M. (2001). Magnetic fabric and inferred
 1446 flow direction of dikes, conesheets and sill swarms, Isle of Skye, Scotland. *Journal of Volcanology*
 1447 *and Geothermal Research* **106**, 195-210.

1448 Hickey, J., Gottsmann, J., Nakamichi, H., Iguchi, M. (2016). Thermomechanical controls on magma
 1449 supply and volcanic deformation: application to Aira caldera, Japan. *Scientific Reports* **6**, 32691.
 1450 Hickey, J., Gottsmann, J., Mothes, P., Odbert, H., Prutkin, I., Vajda, P. (2017). The ups and downs
 1451 of volcanic unrest: insights from integrated geodesy and numerical modelling. *Advances in*
 1452 *Volcanology*, 1-17, Springer.
 1453 Hildreth, W. (2004). Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono
 1454 Craters: several contiguous but discrete systems. *Journal of Volcanology and Geothermal Research*
 1455 **136**, 169-198.
 1456 Hodgetts, D. (2013). Laser scanning and digital outcrop geology in the petroleum industry: A
 1457 review. *Marine and Petroleum Geology* **46**, 335-354.
 1458 Hooft, E. E. E., Morgan, J. V., Nomikou, P., Toomey, D. R., Papazachos, C. V., Warner, M., Heath,
 1459 B., Christopoulou, M-E., Lampridou, D., Kementzetzidou, D. (2016). The PROTEUS Experiment:
 1460 Active Source Seismic Imaging of the Crustal Magma Plumbing Structure of the Santorini Arc
 1461 Volcano. *AGU Fall meeting*, Abstract# DI23B-2619.
 1462 Holohan, E. P., Sudhaus, H., Walter, T. R., Schöpfer, M. P., Walsh, J. J. (2017). Effects of Host-
 1463 rock Fracturing on Elastic-deformation Source Models of Volcano Deflation. *Scientific Reports* **7**,
 1464 10970.
 1465 Hrouda, F. (1982). Magnetic anisotropy of rocks and its application in geology and geophysics.
 1466 *Geophysical Surveys* **5**, 37-82.
 1467 Hrouda, F., Schulmann, K., Suppes, M., Ullemayer, K., de Wall, H., Weber, K. (1997). Quantitative
 1468 relationship between low-field AMS and phyllosilicate fabric: a review. *Physics and Chemistry of*
 1469 *the Earth* **22**, 153-156.
 1470 Hübert, J., Whaler, K. A., Fisseha, S. (2018). The electrical structure of the Central Main Ethiopian
 1471 Rift as imaged by Magnetotellurics - Implications for magma storage and pathways. *Journal of*
 1472 *Geophysical Research*, in review.

1473 Hunt, T. & Bowyer, D. (2007). Reinjection and gravity changes at Rotokawa geothermal field, New
1474 Zealand. *Geothermics* **36**, 421-435.

1475 Husen, S., Taylor, R., Smith, R. B., Healsen, H. (2004). Changes in geyser eruption behavior and
1476 remotely triggered seismicity in Yellowstone National Park produced by the 2002 M 7.9 Denali
1477 fault earthquake, Alaska. *Geology* **32**(6), 537-540.

1478 Ivanic, T., Zibra, I., Doublier, M., Wyche, S. (2013). Geological Interpretation of the Youanmi and
1479 Southern Carnarvon seismic lines 10GA-YU1, 10GA-YU2, 10GA-YU3, and 11GA-SC1. In:
1480 Wyche, S., Ivanic, T., Zibra, I. (eds.) *Youanmi and Southern Carnarvon seismic and*
1481 *magnetotelluric (MT) workshop 2013*. Geological Survey of Western Australia, Plate 3.

1482 Jachens, R. C. & Roberts, C. W. (1985). Temporal and areal gravity investigations at Long Valley
1483 Caldera, California. *Journal of Geophysical Research* **90**, 11210-11218.

1484 Jackson, C. A.-L., Schofield, N., Golenkov, B. (2013). Geometry and controls on the development
1485 of igneous sill-related forced folds: A 2-D seismic reflection case study from offshore southern
1486 Australia. *Geological Society of America Bulletin* **125**, 1874-1890.

1487 Jackson, I., Fitzgerald, J. D., Faul, U. H., Tan, B. H. (2002). Grain-size-sensitive seismic wave
1488 attenuation in polycrystalline olivine. *Journal of Geophysical Research* **107**, B12, 2156-2202.

1489 Jaxybulatov, K., Shapiro, N. M., Koulakov, I., Mordret, A., Landes, M., Sens-Schonfelder, C. (2014).
1490 A large magmatic sill complex beneath the Toba caldera. *Science* **346**, 617-619.

1491 Jay, J. A., Pritchard, M. E., West, M. E., Christensen, D. H., Haney, M., Minaya, E., Sunagua, M.,
1492 McNutt, S. R., Zabala, M. (2012). Shallow seismicity, triggered seismicity, and ambient noise
1493 tomography at the long-dormant Uturuncu volcano, Bolivia. *Bulletin of Volcanology* **74**, 817-837.

1494 Jay, J., Costa, F., Pritchard, M., Lara, L., Singer, B., Herrin, J. (2014). Locating magma reservoirs
1495 using InSAR and petrology before and during the 2011–2012 Cordón Caulle silicic eruption. *Earth*
1496 *and Planetary Science Letters* **395**, 254-266.

1497 Johnson, J. H. & Poland, M. P. (2013). Seismic detection of increased degassing before Kīlauea's
1498 2008 summit explosion. *Nature communications* **4**, 1668.

1499 Johnson, N. E., Whaler, K. A., Hautot, S., Fisseha, S., Desissa, M. & Dawes, G. J. K. (2016). Magma
1500 imaged magnetotellurically beneath an active and an inactive magmatic segment in Afar, Ethiopia.
1501 *Geological Society, London, Special Publications* **420**, 105-125.

1502 Jones, A. G. (2012). Distortion of magnetotelluric data: its identification and removal, in *The*
1503 *Magnetotelluric Method: Theory and Practice*, ed. Chave, A. D. & Jones, A. G., Cambridge
1504 University Press.

1505 Jousset, P., Mori, H. & Okada, H. (2000). Possible magma intrusion revealed by temporal gravity,
1506 ground deformation and ground temperature observations at Mount Komagatake (Hokkaido) during
1507 the 1996-1998 crisis. *Geophysical Journal International* **143**, 557-574.

1508 Kalinicheva, T., Warner, M., Ashley, J., Mancini, F. (2017). Two vs three-dimensional FWI in a 3D
1509 world. *SEG International Exhibition and 87th Annual Meeting*.

1510 Kamei, R., Jang, U. G., Lumley, D., Mouri, T., Nakatsukasa, M., Kato, A. & Takanashi, M. (2017).
1511 Time-lapse Full Waveform Inversion for Monitoring Near-surface Microbubble Injection. *79th*
1512 *Conference and Exhibition, EAGE, Extended Abstracts*.

1513 Kapoor, S., Vigh, D., Wiarda, E., Alwon, S. (2013). Full waveform inversion around the world. *75th*
1514 *Conference and Exhibition, EAGE, Extended Abstracts*.

1515 Karato, S. & Jung, H. (1998). Water, partial melting and the origin of the seismic low velocity and
1516 high attenuation zone in the upper mantle. *Earth and Planetary Science Letters* **157**, 193-207.

1517 Kavanagh, J. L., Burns, A. J., Hazim, S. H., Wood, E., Martin, S. A., Hignett, S. Dennis, D. J. 2018.
1518 Challenging dyke ascent models using novel laboratory experiments: Implications for reinterpreting
1519 evidence of magma ascent and volcanism. *Journal of Volcanology and Geothermal Research*.

1520 Keir, D., Hamling, I. J., Ayele, A., Calais, E., Ebinger, C. J., Wright, T. J., Jacques, E., Mohamed,
1521 K., Hammond, J. O. S., Belachew, M., Baker, E., Rowland, J. V., Lewi, E., Bennati, L. (2009).
1522 Evidence for focused magmatic accretion at segment centers from lateral dike injection captured
1523 beneath the Red Sea rift of Afar. *Geology* **37**, 59-62.

1524 Keir, D., Belachew, M., Ebinger, C. J., Kendall, J. -M., Hammond, J. O. S., Stuart, G. W., Ayele, A.,

1525 Rowland, J. R. (2011). Mapping the evolving strain field during continental breakup from crustal
 1526 anisotropy in the Afar Depression. *Nature Communications* **2**, 285–7.

1527 Kelso, P. R., Tikoff, B., Jackson, M., Sun, W. (2002). A new method for the separation of
 1528 paramagnetic and ferromagnetic susceptibility anisotropy using low field and high field methods.
 1529 *Geophysical Journal International* **151**, 345-359.

1530 Kent, G. M., Singh, S. C., Harding, A. J., Sinha, M. C., Orcutt, J. A. (2000). Evidence from three-
 1531 dimensional seismic reflectivity images for enhanced melt supply beneath mid-ocean-ridge
 1532 discontinuities. *Nature* **406**, 614-618.

1533 Key, K., Constable, S., Liu, L., Pommier, A. (2013). Electrical image of passive upwelling beneath
 1534 the northern East Pacific Rise. *Nature* **495**, 499-502.

1535 Khan, M. A. (1962). The anisotropy of magnetic susceptibility of some igneous and metamorphic
 1536 rocks. *Journal of Geophysical Research* **67**, 2873-2885.

1537 Kim, D., Brown, L. D., Árnason, K., Ágústsson, K., Blanck, H. (2017). Magma reflection imaging in
 1538 Krafla, Iceland, using microearthquake sources. *Journal of Geophysical Research Solid Earth* **122**,
 1539 5228-5242.

1540 Kiser, E., Levander, A., Schmandt, B., Palomeras, I., Harder, S. H., Creager, K. C., Vidale, J. E.,
 1541 Malone, S. D. (2014). Field report on the iMUSH active source seismic experiment. *AGU Fall*
 1542 *Meeting Abstracts* **1**, 2.

1543 Knight, M. D. & Walker, G. P. (1988). Magma flow directions in dikes of the Koolau Complex,
 1544 Oahu, determined from magnetic fabric studies. *Journal of Geophysical Research* **93**, 4301-4319.

1545 Koulakov, I., Gordeev, E. I., Dobretsov, N. L., Vernikovskiy, V. A., Senyukov, S., Jakovlev, A. &
 1546 Jaxybulatov, K. (2013). Rapid changes in magma storage beneath the Klyuchevskoy group of
 1547 volcanoes inferred from time-dependent seismic tomography. *Journal of Volcanology and*
 1548 *Geothermal Research* **263**, 75-91.

1549 Laumonier, M., Gaillard, F., Sifre, D. (2015). The effect of pressure and water concentration on the
 1550 electrical conductivity of dacitic melts: Implication for magnetotelluric imaging in subduction areas.
 1551 *Chemical Geology* **418**, 66-76.

1552 Launeau, P. & Cruden, A. R. (1998). Magmatic fabric acquisition mechanisms in a syenite: Results
 1553 of a combined anisotropy of magnetic susceptibility and image analysis study. *Journal of*
 1554 *Geophysical Research* **103**, 5067-5089.

1555 Lazaratos, S., Chikichev, I., Wang, K. (2011). Improving the convergence rate of full wavefield
 1556 inversion using spectral shaping. *81st Annual International Meeting, SEG, Expanded Abstracts*.

1557 Le Mével, H., Feigl, K. L., Córdova, L., DeMets, C. & Lundgren, P. (2015). Evolution of unrest at
 1558 Laguna del Maule volcanic field (Chile) from InSAR and GPS measurements, 2003 to 2014.
 1559 *Geophysical Research Letters* **42**, 6590-6598.

1560 Leberl, F., Irschara, A., Pock, T., Meixner, P., Gruber, M., Scholz, S., Wiechert, A. (2010). Point
 1561 Clouds: Lidar versus 3D Vision. *Photogrammetric Engineering and Remote Sensing* **76**, 1123-
 1562 1134.

1563 Lees, J. M. (2007). Seismic tomography of magmatic systems. *Journal of Volcanology and*
 1564 *Geothermal Research* **167**, 37-56.

1565 Lu, Z. Dzurisin, D. (2014). InSAR Imaging of Aleutian Volcanoes. In *InSAR Imaging of Aleutian*
 1566 *Volcanoes* (pp. 87-345). Springer Berlin Heidelberg.

1567 MacGregor, L. M., Constable, S., Sinha, M. C. (1998). The RAMESSES experiment—III.
 1568 Controlled-source electromagnetic sounding of the Reykjanes Ridge at 57° 45' N. *Geophysical*
 1569 *Journal International* **135**, 773-789.

1570 Magee, C., Stevenson, C. T. E., O'Driscoll, B., Schofield, N., McDermott, K. (2012a). An
 1571 alternative emplacement model for the classic Ardnamurchan cone sheet swarm, NW Scotland,
 1572 involving lateral magma supply via regional dykes. *Journal of Structural Geology* **43**, 73-91.

1573 Magee, C., Stevenson, C. T. E., O'Driscoll, B., Petronis, M. S. (2012b). Local and regional controls
1574 on the lateral emplacement of the Ben Hiant Dolerite intrusion, Ardnamurchan (NW Scotland).
1575 *Journal of Structural Geology* **39**, 66-82.

1576 Magee, C., Hunt-Stewart, E., Jackson, C. A. L. (2013a). Volcano growth mechanisms and the role
1577 of sub-volcanic intrusions: Insights from 2D seismic reflection data. *Earth and Planetary Science*
1578 *Letters* **373**, 41-53.

1579 Magee, C., Briggs, F., Jackson, C. A.-L. (2013b). Lithological controls on igneous intrusion-
1580 induced ground deformation. *Journal of the Geological Society* **170**, 853-856.

1581 Magee, C., Jackson, C. L., Schofield, N. (2014). Diachronous sub-volcanic intrusion along deep-
1582 water margins: insights from the Irish Rockall Basin. *Basin Research* **26**, 85-105.

1583 Magee, C., Maharaj, S. M., Wrona, T., Jackson, C. A.-L. (2015). Controls on the expression of
1584 igneous intrusions in seismic reflection data. *Geosphere*, GES01150.01151.

1585 Magee, C., Muirhead, J. D., Karvelas, A., Holford, S. P., Jackson, C. A., Bastow, I. D., Schofield,
1586 N., Stevenson, C. T., McLean, C., McCarthy, W. (2016). Lateral magma flow in mafic sill
1587 complexes. *Geosphere*, GES01256.01251.

1588 Magee, C., Bastow, I. D., van Wyk de Vries, B., Jackson, C. A. L., Hetherington, R., Hagos, M. &
1589 Hoggett, M. (2017). Structure and dynamics of surface uplift induced by incremental sill
1590 emplacement. *Geology* **45**, 431-434.

1591 Magee, C., Muirhead, J., Schofield, N., Walker, R., Galland, O., Holford, S., Spacapan, J., Jackson,
1592 C. A.-L., McCarthy, W. (2017). Structural signatures of igneous sheet intrusion propagation.
1593 *EarthArXiv* doi:10.17605/OSF.IO/PDN42.

1594 Malehmir, A., Durrheim, R., Bellefleur, G., Urosevic, M., Juhlin, C., White, D. J., Milkereit, B.,
1595 Campbell, G. (2012). Seismic methods in mineral exploration and mine planning: A general
1596 overview of past and present case histories and a look into the future. *Geophysics* **77**, WC173-
1597 WC190.

1598 Mamtani, M. A., Pal, T. & Greiling, R. O. (2013). Kinematic analysis using AMS data from a
1599 deformed granitoid. *Journal of Structural Geology* **50**, 119-132.

1600 Mancini, F., Moss, J., Prindle, K., Ridsdill-Smith, T. (2015). Where can full waveform inversion
1601 have the biggest impact in the exploration and production cycle? *77th International Conference and*
1602 *Exhibition, EAGE, Extended Abstracts*.

1603 Manzella, A., Volpi, G., Zaja, A., Meju, M. (2004). Combined TEM-MT investigation of shallow-
1604 depth resistivity structure of Mt Somma-Vesuvius. *Journal of Volcanology and Geothermal*
1605 *Research* **131**, 19-32.

1606 Marjanovic, M., Carbotte, S. M., Carton, H., Nedimovic, M. R., Mutter, J. C., Canales, J. P. (2014).
1607 A multi-sill magma plumbing system beneath the axis of the East Pacific Rise. *Nature Geoscience* **7**,
1608 825-829.

1609 Mark, N. J., Schofield, N., Pugliese, S., Watson, D., Holford, S., Muirhead, D., Brown, R., Healy, D.
1610 (2017). Igneous intrusions in the Faroe Shetland basin and their implications for hydrocarbon
1611 exploration; new insights from well and seismic data. *Marine and Petroleum Geology*

1612 Masterlark, T. (2007). Magma intrusion and deformation predictions: Sensitivities to the Mogi
1613 assumptions. *Journal of Geophysical Research: Solid Earth*, **112**(B6).

1614 Masturyono., McCaffrey, R., Wark, D. A., Roecker, S. W., Fauzi Ibrahim, G. & Sukhyar. (2001).
1615 Distribution of magma beneath Toba caldera complex, north Sumatra, Indonesia, constrained by
1616 three-dimensional P wave velocities, seismicity, and gravity. *Geochemistry Geophysics Geosystems*
1617 **2**, 1014.

1618 Matsushima, N., Oshima, H., Ogawa, Y., Takakura, S., Satoh, H., Utsugi, M., Nishida, Y. (2011).
1619 Magma prospecting in Usu volcano, Hokkaido, Japan, using magnetotelluric soundings. *Journal of*
1620 *Volcanology and Geothermal Research* **109**, 263-277.

1621 McBride, J. H., William Keach, R., Leetaru, H. E. & Smith, K. M. (2018). Visualizing Precambrian
1622 basement tectonics beneath a carbon capture and storage site, Illinois Basin. *Interpretation* **6**, T257-
1623 T270.

1624 McCabe, C., Jackson, M., Ellwood, B. B. (1985). Magnetic anisotropy in the Trenton limestone:
 1625 results of a new technique, anisotropy of anhysteretic susceptibility. *Geophysical Research Letters*
 1626 **12**, 333-336.

1627 McCormick Kilbride, B., Edmonds, M. & Biggs, J. (2016). Observing eruptions of gas-rich
 1628 compressible magmas from space. *Nature communications* **7**, 13744.

1629 McKenzie, D. A. N., & O'Nions, R. K. (1991). Partial melt distributions from inversion of rare earth
 1630 element concentrations. *Journal of Petrology* **32**, 1021-1091.

1631 McLean, C. E., Schofield, N., Brown, D. J., Jolley, D. W., Reid, A. (2017). 3D seismic imaging of
 1632 the shallow plumbing system beneath the Ben Nevis Monogenetic Volcanic Field: Faroe-Shetland
 1633 Basin. *Journal of the Geological Society* **174**, 468-485.

1634 Miller, V. & Savage, M. (2001). Changes in seismic anisotropy after volcanic eruptions: evidence
 1635 from Mount Ruapehu. *Science* **293**, 2231-2233.

1636 Miller, C. F., Furbish, D. J., Walker, B. A., Claiborne, L. L., Koteas, G. C., Bleick, H. A., Miller, J.
 1637 S. (2011). Growth of plutons by incremental emplacement of sheets in crystal-rich host: evidence
 1638 from Miocene intrusions of the Colorado River region, Nevada, USA. *Tectonophysics* **500**, 65-77.

1639 Miller, K. J., Zhu, W., Montesi, L. G. J., Gaetani, G. A. (2014). Experimental quantification of
 1640 permeability of partially molten mantle rock. *Earth and Planetary Science Letters* **388**, 273-282.

1641 Miller, C. A., Le Mével, H., Currenti, G., Williams-Jones, G. & Tikoff, B. (2017). Microgravity
 1642 changes at the Laguna del Maule volcanic field: Magma-induced stress changes facilitate mass
 1643 addition. *Journal of Geophysical Research: Solid Earth* **122**, 3179-3196.

1644 Mitchell, M. A., White, R. S., Roecker, S., Greenfield, T. (2013). Tomographic image of melt
 1645 storage beneath Askja Volcano, Iceland using local microseismicity. *Geophysical Research Letters*
 1646 **40**, 5040-5046.

1647 Morgan, J. V., Warner, M., Bell, R. E., Ashley, J., Barnes, D., Little, R., Roele, K., Jones, C.
 1648 (2013). Next-generation seismic experiments: wide-angle, multi-azimuth, three-dimensional, full-
 1649 waveform inversion. *Geophysical Journal International* **195**, 1657-1678.

1650 Morgan, J. V., Warner, M., Arnoux, G., Hooft, E., Toomey, D., VanderBeek, B., Wilcock, W.
1651 (2016), Next-generation seismic experiments – II: wide-angle, multi-azimuth, 3-D, full-waveform
1652 inversion of sparse field data. *Geophysical Journal International* **204**, 1342-1363.

1653 Muirhead, J. D., Airoidi, G., White, J. D., Rowland, J. V. (2014). Cracking the lid: Sill-fed dikes are
1654 the likely feeders of flood basalt eruptions. *Earth and Planetary Science Letters* **406**, 187-197.

1655 Neuberg, J., Baptie, B., Luckett, R., Stewart, R. (1998). Results from the broadband seismic
1656 network on Montserrat. *Geophysical Research Letters* **25(19)**, 3661-3664.

1657 Neuberg, J., Tuffen, H., Collier, L., Green, D., Powell, T., Dingwell, D. (2006). The trigger
1658 mechanism of low-frequency earthquakes on Montserrat. *Journal of Volcanology and Geothermal*
1659 *Research* **153**, 37-50.

1660 Neves, S. P., Araújo, A. M., Correia, P. B. & Mariano, G. (2003). Magnetic fabrics in the Cabanas
1661 Granite (NE Brazil): interplay between emplacement and regional fabrics in a dextral transpressive
1662 regime. *Journal of Structural Geology* **25**, 441-453.

1663 Newman, A. V., Dixon, T. H., Gourmelen, N. (2006). A four-dimensional viscoelastic deformation
1664 model for Long Valley Caldera, California, between 1995 and 2000. *Journal of Volcanology and*
1665 *Geothermal Research* **150(1)**, 244-269.

1666 O'Driscoll, B., Stevenson, C. T. E., Troll, V. R. (2008). Mineral lamination development in layered
1667 gabbros of the British Palaeogene Igneous Province: A combined anisotropy of magnetic
1668 susceptibility, textural and mineral chemistry study. *Journal of Petrology* **49**, 1187-1221.

1669 O'Driscoll, B., Ferré, E. C., Stevenson, C. T. E., Magee, C. (2015). The significance of
1670 magnetic fabrics in layered mafic-ultramafic intrusions. In: Charlier, B., Latypov, R., Namur,
1671 O., Tegner, C. (Eds), *Layered Intrusions*, Springer, 295-329.

1672 Okubo, P. G., Benz, H. M., Chouet, B. A. (1997). Imaging the crustal magma sources beneath
1673 Mauna Loa and Kilauea volcanoes, Hawaii. *Geology* **25(10)**, 867-870.

1674 Orlický, O. (1990). Detection of magnetic carriers in rocks: results of susceptibility changes in
 1675 powdered rock samples induced by temperature. *Physics of the Earth and Planetary Interiors*
 1676 **63**, 66-70.

1677 Padilha, A. L., Vitorello, Í., Pádua, M. B., Bologna, M. S. (2006). Lithospheric and
 1678 sublithospheric anisotropy beneath central-southeastern Brazil constrained by long period
 1679 magnetotelluric data. *Physics of the Earth and Planetary Interiors* **158**, 190-209.

1680 Pagli, C., Wright, T. J., Ebinger, C. J., Yun, S. H., Cann, J. R., Barnie, T. & Ayele, A. (2012).
 1681 Shallow axial magma chamber at the slow-spreading Erta Ale Ridge. *Nature Geoscience* **5**, 284-
 1682 288.

1683 Parker, R. L. (1980). The inverse problem of electromagnetic induction: Existence and construction
 1684 of solutions based on incomplete data. *Journal of Geophysical Research* **85**, 4421-4428.

1685 Parks, M. M., Biggs, J., England, P., Mather, T. A., Nomikou, P., Palamartchouk, K., Papanikolaou,
 1686 X., Paradissis, D., Parsons, B., Pyle, D. M., Raptakis, C. (2012). Evolution of Santorini Volcano
 1687 dominated by episodic and rapid fluxes of melt from depth. *Nature Geoscience* **5**(10), 749-754.

1688 Parker, A. L. (2016). InSAR Observations of Ground Deformation: Application to the Cascades
 1689 Volcanic Arc. Springer Theses.

1690 Parmigiani, A., Faroughi, S. A., Huber, C., Bachmann, O., Su, Y. (2016) Bubble accumulation and
 1691 its role on the evolution of upper crustal magma reservoirs. *Nature* **532**, 492-495.

1692 Paulatto, M., Minshull, T., Baptie, B., Dean, S., Hammond, J. O. S., Henstock, T., Kenedi, C.,
 1693 Kiddle, E., Malin, P., Peirce, C., Ryan, G. (2010). Upper crustal structure of an active volcano from
 1694 refraction/reflection tomography, Montserrat, Lesser Antilles. *Geophysical Journal International*
 1695 **180**, 685-696.

1696 Paulatto, M., Annen, C., Henstock, T. J., Kiddle, E., Minshull, T. A., Sparks, R. S. J., Voight, B.
 1697 (2012). Magma chamber properties from integrated seismic tomography and thermal modeling at
 1698 Montserrat. *Geochemistry, Geophysics, Geosystems* **13**(1).

1699 Pedersen, R. & Sigmundsson, F. (2006). Temporal development of the 1999 intrusive episode in the
 1700 Eyjafjallajökull volcano, Iceland, derived from InSAR images. *Bulletin of Volcanology* **68**(4), 377-
 1701 393.

1702 Peron-Pinvidic, G., Shillington, D. J., Tucholke, B. E. (2010). Characterization of sills associated
 1703 with the U reflection on the Newfoundland margin: evidence for widespread early post-rift
 1704 magmatism on a magma-poor rifted margin. *Geophysical Journal International* **182**, 113-136.

1705 Petford, N., Cruden, A. R., McCaffrey, K. J. W. & Vigneresse, J. L. (2000). Granite magma
 1706 formation, transport and emplacement in the Earth's crust. *Nature* **408**, 669-673.

1707 Petronis, M. S., Delcamp, A., de Vries, B. V. W. (2013). Magma emplacement into the Lemptégy
 1708 scoria cone (Chaîne Des Puys, France) explored with structural, anisotropy of magnetic
 1709 susceptibility, and Paleomagnetic data. *Bulletin of Volcanology* **75**, 753.

1710 Petronis, M. S., Brister, A. R., Rapprich, V., de Vries, B. V. W., Lindline, J., Misurec, J. (2015).
 1711 Emplacement history of the Trosky basanitic volcano (Czech Republic): paleomagnetic, rock
 1712 magnetic, petrologic, and anisotropy of magnetic susceptibility evidence for lingering growth of a
 1713 monogenetic volcano. *Journal of Geosciences* **60**, 129-147.

1714 Petrovský, E. & Kapička, A. (2006). On determination of the Curie point from thermomagnetic
 1715 curves. *Journal of Geophysical Research* **111**, B12S27.

1716 Phillips, T. B., Magee, C., Jackson, C. A. L. & Bell, R. E. (2017). Determining the three-
 1717 dimensional geometry of a dike swarm and its impact on later rift geometry using seismic reflection
 1718 data. *Geology* **46**, 119-122.

1719 Pinel, V., Poland, M. P., Hooper, A. (2014). Volcanology: Lessons learned from synthetic aperture
 1720 radar imagery. *Journal of Volcanology and Geothermal Research* **289**, 81-113.

1721 Planke, S., Symonds, P. A., Alvestad, E., Skogseid, J. (2000). Seismic volcanostratigraphy of large-
 1722 volume basaltic extrusive complexes on rifted margins. *Journal of Geophysical Research: Solid*
 1723 *Earth* **105**, 19335-19351.

1724 Planke, S., Rasmussen, T., Rey, S. S., Myklebust, R. (2005). Seismic characteristics and
1725 distribution of volcanic intrusions and hydrothermal vent complexes in the Vøring and Møre basins.
1726 In: Doré, A. G. (ed.) *Petroleum Geology: North-West Europe and Global Perspectives* -
1727 *Proceedings of the 6th Petroleum Geology Conference*: Geological Society, London, 833-844.

1728 Poland, M. P., Miklius, A., Sutton, A. J. and Thornber, C. R. (2012). A mantle-driven surge in
1729 magma supply to Kilauea Volcano during 2003-2007. *Nature Geoscience* **5**(4), 295-300.

1730 Poland, M. P. & Carbone, D. (2016). Insights into shallow magmatic processes at Kīlauea Volcano,
1731 Hawai‘i, from a multiyear continuous gravity time series. *Journal of Geophysical Research: Solid*
1732 *Earth* **121**, 5477-5492.

1733 Pommier, A. (2014). Interpretation of magnetotelluric results using laboratory measurements.
1734 *Surveys in Geophysics* **35**, 41-84.

1735 Pommier, A., Tarits, P., Hautot, S., Pichavant, M., Scaillet, B. & Gaillard, F. (2010). A new
1736 petrological and geophysical investigation of the present-day plumbing system of Mount
1737 Vesuvius. *Geochemistry Geophysics Geosystems* **11**, Q07013.

1738 Pommier, A. & Le Trong, E. (2011). SIGMELTS: A web portal for electrical conductivity
1739 calculations in geosciences. *Computational Geoscience* **37**, 1450-1459.

1740 Pritchard, M. E. & Simons, M. (2004). An InSAR-based survey of volcanic deformation in the
1741 central Andes. *Geochemistry, Geophysics, Geosystems* **5**(2).

1742 Prutkin, I., Vajda, P. & Gottsmann, J. (2014). The gravimetric picture of magmatic and
1743 hydrothermal sources driving hybrid unrest on Tenerife in 2004/5. *Journal of Volcanology and*
1744 *Geothermal Research* **282**, 9-18.

1745 Reeves, J., Magee, C. & Jackson, C. A. L. (2018). Unravelling intrusion-induced forced fold
1746 kinematics and ground deformation using 3D seismic reflection data. *Volcanica* **1**, 1-17.

1747 Reich, H. (1932). Die Bedeutung der finnischen Schweremessungen für die angewandte Geophysik:
1748 Beitr. Angew. Geophys., *Erganzungsh* **2**, 1-13.

1749 Richter, C. & van der Pluijm, B.A. (1994). Separation of paramagnetic and ferrimagnetic
1750 susceptibilities using low temperature magnetic susceptibilities and comparison with high field
1751 methods. *Physics of the Earth and Planetary Interiors* **82**, 113-123.

1752 Rivalta, E. & Segall, P. (2008). Magma compressibility and the missing source for some dike
1753 intrusions. *Geophysical Research Letters* **35**(4).

1754 Roberts, J. J. & Tyburczy, J. A. (1999). Partial-melt electrical conductivity: Influence of melt
1755 composition. *Journal of Geophysical Research* **104**, 7055-7065.

1756 Rocchi, S., Mazzotti, A., Marroni, M., Pandolfi, L., Costantini, P., Giuseppe, B., Biase, D. D.,
1757 Federici, F. & Lo, P. G. (2007). Detection of Miocene saucer-shaped sills (offshore Senegal) via
1758 integrated interpretation of seismic, magnetic and gravity data. *Terra Nova* **19**, 232-239.

1759 Rochette, P., Jackson, M., Aubourg, C. (1992). Rock magnetism and the interpretation of
1760 anisotropy of magnetic susceptibility. *Reviews of Geophysics* **30**, 209-226.

1761 Roman, D. C. & Cashman, K.V. (2006). The origin of volcano-tectonic earthquake swarms.
1762 *Geology* **34**, 457-460.

1763 Roman, D. C., Savage, M. K., Arnold, R., Latchman, J. L. & De Angelis, S. (2011). Analysis and
1764 forward modeling of seismic anisotropy during the ongoing eruption of the Soufrière Hills Volcano,
1765 Montserrat, 1996–2007. *Journal of Geophysical Research: Solid Earth* **116**, B03201.

1766 Ronchin, E., Masterlark, T., Dawson, J., Saunders, S., Martí Molist, J. (2017). Imaging the complex
1767 geometry of a magma reservoir using FEM-based linear inverse modeling of InSAR data:
1768 application to Rabaul Caldera, Papua New Guinea. *Geophysical Journal International* **209**, 1746-
1769 1760.

1770 Routh, P., Neelamani, R., Lu, R., Lazaratos, S., Braaksma, H., Hughes, S., Saltzer, R., Stewart, J.,
1771 Naidu, K., Averill, H. & Gottumukkula, V. (2017). Impact of high-resolution FWI in the western
1772 black sea: Revealing overburden and reservoir complexity. *The Leading Edge* **36**, 60-66.

1773 Rubin, A. & Gillard, D. (1998). Dike-induced earthquakes: Theoretical considerations. *Journal of*
1774 *Geophysical Research* **103**, 10017-10030.

1775 Rymer, H. & Brown, G. C. (1986). Gravity fields and the interpretation of volcanic structures:
 1776 Geological discrimination and temporal evolution. *Journal of Volcanology and Geothermal*
 1777 *Research* **27**, 229-254.

1778 Rymer, H., Locke, C. A., Brenes, J. & Williams-Jones, G. (2005). Magma plumbing processes for
 1779 persistent activity at Poas volcano, Costa Rica. *Geophysical Research Letters* **32**, L08307.

1780 Salzer, J. T., Nikkhoo, M., Walter, T. R., Sudhaus, H., Reyes-Dávila, G., Bretón, M., Arámbula, R.
 1781 (2014). Satellite radar data reveal short-term pre-explosive displacements and a complex conduit
 1782 system at Volcán de Colima, Mexico. *Frontiers in Earth Science* **2**, 12.

1783 Samrock, F., Kuvshinov, A., Bakker, J., Jackson, A., Fisseha, S. (2015). 3D analysis and
 1784 interpretation of Magnetotelluric data from the Aluto-Langano geothermal field, Ethiopia.
 1785 *Geophysical Journal International* **202**, 1923-1948

1786 Saunders, K., Blundy, J. D., Dohmen, R., Cashman, K. (2012). Linking Petrology and Seismology
 1787 at an active volcano. *Science* **336**, 1023-1027.

1788 Saxby, J., Gottsmann, J., Cashman, K. & Gutiérrez, E. (2016). Magma storage in a strike-slip
 1789 caldera. *Nature Communications* **7**, 12295.

1790 Schaefer, L. N., Lu, Z., Oommen, T. (2015). Dramatic volcanic instability revealed by InSAR.
 1791 *Geology* **43**, 743-746.

1792 Schmeling, H. (1986). Numerical models on the influence of partial melt on elastic, anelastic and
 1793 electrical properties of rocks. Part II: electrical conductivity. *Physics of Earth and Planetary*
 1794 *Interiors* **43**, 123-136.

1795 Schofield, N., Stevenson, C., Reston, T. (2010). Magma fingers and host rock fluidization in the
 1796 emplacement of sills. *Geology* **38**, 63-66.

1797 Schofield, N., Heaton, L., Holford, S. P., Archer, S. G., Jackson, C. A. L., Jolley, D. W. (2012a).
 1798 Seismic imaging of 'broken bridges': linking seismic to outcrop-scale investigations of intrusive
 1799 magma lobes. *Journal of the Geological Society of London* **169**, 421-426.

1800 Schofield, N. J., Brown, D. J., Magee, C., Stevenson, C. T. E. (2012b) Sill morphology and
 1801 comparison of brittle and non-brittle emplacement mechanisms. *Journal of the Geological Society*
 1802 *of London* **169**, 127-141.

1803 Schofield, N., Holford, S., Millett, J., Brown, D., Jolley, D., Passey, S., Muirhead, D., Grove, C.,
 1804 Magee, C., Murray, J., Hole, M., Jackson, C., Stevenson, C. (2017). Regional Magma Plumbing and
 1805 emplacement mechanisms of the Faroe-Shetland Sill Complex: Implications for magma transport
 1806 and petroleum systems within sedimentary basins. *Basin Research* **29**, 41-63.

1807 Segall, P. (2010). *Earthquake and volcano deformation*. Princeton University Press.

1808 Shalev, E., Kenedi, C. L., Malin, P., Voight, V., Miller, V., Hidayat, D., Sparks, R. S. J., Minshull,
 1809 T., Paulatto, M., Brown, L. & Mattioli, G. (2010). Three-dimensional seismic velocity tomography
 1810 of Montserrat from the SEA-CALIPSO offshore/onshore experiment. *Geophysical Research Letters*
 1811 **37**, L00E17.

1812 Sigmundsson, F., Hreinsdóttir, S., Hooper, A., Árnadóttir, T., Pedersen, R., Roberts, M.J.,
 1813 Óskarsson, N., Auriac, A., Decriem, J., Einarsson, P., Geirsson, H. (2010). Intrusion triggering of
 1814 the 2010 Eyjafjallajökull explosive eruption. *Nature* **468**(7322), 426-430.

1815 Sigmundsson, F. & 37 others (2015). Segmented lateral dyke growth in a rifting event at
 1816 Bardabunga volcanic system, Iceland. *Nature* **517**, 191-195.

1817 Sigmundsson, F. (2016), New insights into magma plumbing along rift systems from detailed
 1818 observations of eruptive behavior at Axial volcano. *Geophysical Research Letters* **43**, 12,423-
 1819 12,427.

1820 Simons, M. & P. A. Rosen. (2007). Interferometric synthetic aperture radar geodesy. *Treatise on*
 1821 *Geophysics - Geodesy* **3**, 391-446, Elsevier, Amsterdam.

1822 Simpson, F. & Bahr, K. (2005). *Practical Magnetotellurics*. Cambridge University Press,
 1823 Cambridge, UK.

1824 Singer, B. S., Andersen, N. L., Le Mével, H., Feigl, K. L., DeMets, C., Tikoff, B., Thurber, C. H.,
 1825 Jicha, B. R., Cardona, C., Córdova, L. and Gil, F. (2014). Dynamics of a large, restless, rhyolitic
 1826 magma system at Laguna del Maule, southern Andes, Chile. *GSA Today* **24**, 4-10.
 1827 Sirgue, L., Barkved, O. I., Dellinger, J., Etgen, J., Albertin, U., Kommedal, J. H. (2010) Full
 1828 waveform inversion: the next leap forward in imaging at Valhall. *First Break* **28**, 65-70.
 1829 Smallwood, J. R. & Maresh, J. (2002). The properties, morphology and distribution of igneous sills:
 1830 modelling, borehole data and 3D seismic from the Faroe-Shetland area. In: Jolley, D. W. & Bell, B.
 1831 R. (eds.) *The North Atlantic Igneous Province: Stratigraphy, tectonic, Volcanic and Magmatic*
 1832 *Processes*: Geological Society, London, Special Publications, 271-306.
 1833 Sofyan, Y., Kamah, Y., Nishijima, J., Fujimitsu, Y., Ehara, S., Fukuda, Y. & Taniguchi, M. (2011).
 1834 Mass variation in outcome to high production activity in Kamojang Geothermal Field, Indonesia: A
 1835 reservoir monitoring with relative and absolute gravimetry. *Earth, Planets and Space* **63**, 1157-
 1836 1167.
 1837 Solano, J. M. S., Jackson, M. D., Sparks, R. S. J., Blundy, J. D., Annen, C. (2012). Melt segregation
 1838 in deep crustal hot zones: a mechanism for chemical differentiation, crustal assimilation and the
 1839 formation of evolved magmas. *Journal of Petrology* **53**, 1999-2026.
 1840 Sparks, R., Biggs, J., Neuberg, J. (2012). Monitoring volcanoes. *Science* **335**, 1310-1311.
 1841 Stacey, F. D., Joplin, G., Lindsay, J. (1960). Magnetic anisotropy and fabric of some foliated rocks
 1842 from S.E. Australia. *Pure Applied Geophysics* **47**, 30-40.
 1843 Stankiewicz, J., Ryberg, T., Haberland, C., Natawidjaja, D. (2010). Lake Toba volcano magma
 1844 chamber imaged by ambient seismic noise tomography. *Geophysical Research Letters* **37**, L17306.
 1845 Stephenson, A. (1994). Distribution anisotropy: two simple models for magnetic lineation and
 1846 foliation. *Physics of the Earth and Planetary Interiors* **82**, 49-53.
 1847 Sternberg, B., Washburne, J. C., Pellerin, L. (1988). Correction for the static shift in
 1848 magnetotellurics using transient electromagnetic soundings. *Geophysics* **53**, 1459-1468.

1849 Stevenson, C. T. E., Owens, W. H., Hutton, D. H. W. (2007a). Flow lobes in granite: the
1850 determination of magma flow direction in the Trawenagh Bay Granite, N.W. Ireland, using
1851 anisotropy of magnetic susceptibility. *Geological Society of America Bulletin* **119**, 1368-1386.

1852 Stevenson, C. T. E., Owens, W. H., Hutton, D. H. W., Hood, D. N., Meighan, I. G. (2007b).
1853 Laccolithic, as opposed to cauldron subsidence, emplacement of the Eastern Mourne pluton, N.
1854 Ireland: evidence from anisotropy of magnetic susceptibility. *Journal of the Geological Society of*
1855 *London* **164**, 99-110.

1856 Stork, A., Stuart, G. W., Henderson, C. M., Keir, D., Hammond, J. O. S. (2013). Uppermost mantle
1857 (Pn) velocity model for the Afar region, Ethiopia: An insight into rifting processes. *Geophysical*
1858 *Journal International* **193**, 321-328.

1859 Stuart, G. W., Bastow, I. D., Ebinger, C. J. (2006). Crustal structure of the Northern Main Ethiopian
1860 rift from receiver function studies. In: Yirgu, G., Ebinger, C. J., Maguire, P. K. H. (eds) The Afar
1861 Volcanic Province within the East African Rift System. *Geological Society, London, Special*
1862 *Publications* **259**, 253-267.

1863 Sturkell, E., Einarsson, P., Sigmundsson, F., Geirsson, H., Olafsson, H., Pedersen, R., de Zeeuw-
1864 van Dalfsen, E., Linde, A.T., Sacks, S.I., Stefánsson, R. (2006). Volcano geodesy and magma
1865 dynamics in Iceland. *Journal of Volcanology and Geothermal Research* **150**, 14-34.

1866 Svensen, H., Corfu, F., Polteau, S., Hammer, Ø., Planke, S. (2012). Rapid magma emplacement in
1867 the Karoo Large Igneous Province. *Earth and Planetary Science Letters* **325-326**, 1-9.

1868 Symonds, P. A., Planke, S., Frey, O., Skogseid, J. (1998). Volcanic evolution of the Western
1869 Australian Continental Margin and its implications for basin development. *The Sedimentary Basins*
1870 *of Western Australia 2: Proc. of Petroleum Society Australia Symposium, Perth, WA.*

1871 Takada, Y. & Fukushima, Y. (2013). Volcanic subsidence triggered by the 2011 Tohoku earthquake
1872 in Japan. *Nature Geoscience* **6**(8), 637-641.

1873 Takei, Y. (2002). Effect of pore geometry on V-P/V-S: From equilibrium geometry to crack.
1874 *Journal of Geophysical Research-Solid Earth* **107**, 2043.

1875 Tarantola, A. (1984) Inversion of seismic reflection data in the acoustic approximation. *Geophysics*
1876 **49**, 1259-1266.

1877 Tarling, D. H. & Hrouda, F. (1993). The Magnetic Anisotropy of Rocks, Chapman and Hall, New
1878 York, pp. 1-232.

1879 Tauxe, L. (1998). Paleomagnetic Principles and Practice. In: Of Modern approaches in geophysics,
1880 vol. 17: Kluwer Academic Publishers, Dordrecht, Boston, London.

1881 Thiele, S., Micklethwaite, S., Bourke, P., Verrall, M., Kovesi, P. (2015). Insights into the mechanics
1882 of en-echelon sigmoidal vein formation using ultra-high resolution photogrammetry and computed
1883 tomography. *Journal of Structural Geology* **77**, 27-54.

1884 Thomas, M. E. & Neuberg, J. (2012). What makes a volcano tick - A first explanation of deep
1885 multiple seismic sources in ascending magma. *Geology* **40**, 351-354.

1886 Thomson, K. (2007). Determining magma flow in sills, dykes and laccoliths and their implications
1887 for sill emplacement mechanisms. *Bulletin of Volcanology* **70**, 183-201.

1888 Thomson, K. & Hutton, D. (2004). Geometry and growth of sill complexes: insights using 3D
1889 seismic from the North Rockall Trough. *Bulletin of Volcanology* **66**, 364-375.

1890 Tizzani, P., Battaglia, M., Zeni, G., Atzori, S., Berardino, P. & Lanari, R. (2009). Uplift and magma
1891 intrusion at Long Valley caldera from InSAR and gravity measurements. *Geology* **37**, 63-66.

1892 Tolstoy, M., Waldhauser, F., Bohnenstiehl, D. R., Weekly, R. T., Kim, W. -Y. (2008). Seismic
1893 identification of along-axis hydrothermal flow on the East Pacific Rise. *Nature* **451**, 181-184.

1894 Trasatti, E., Giunchi, C. & Agostinetti, N. P. (2008). Numerical inversion of deformation caused by
1895 pressure sources: application to Mount Etna (Italy). *Geophysical Journal International* **172**, 873-
1896 884.

1897 Trindade, R. I. F., Nguema, T. M. M., Bouchez, J. L. (2001). Thermally enhanced mimetic fabric of
1898 magnetite in a biotite granite. *Geophysical Research Letters* **28**, 2687-2690.

1899 Trude, J., Cartwright, J., Davies, R. J. & Smallwood, J. (2003). New technique for dating igneous
1900 sills. *Geology* **31**, 813-816.

1901 Tuffen, H., Smith, R., Sammonds, P. R. (2008). Evidence for seismogenic fracture of silicic
1902 magma. *Nature* **453**, 511-514.

1903 Van Camp, M., de Viron, O., Watlet, A., Meurers, B., Francis, O. & Caudron, C. (2017).
1904 Geophysics from terrestrial time-variable gravity measurements. *Reviews of Geophysics* **55**, 938-
1905 992

1906 Vargas-Bracamontes, D. M. & Neuberg, J. W. (2012). Interaction between regional and magma-
1907 induced stresses and their impact on volcano-tectonic seismicity. *Journal of Volcanology and*
1908 *Geothermal Research* **243**, 91-96.

1909 Vasuki, Y., Holden, E. -J., Kovesi, P., Micklethwaite, S. (2014) Semi-automatic mapping of
1910 geological Structures using UAV-based photogrammetric data: An image analysis approach.
1911 *Computers and Geosciences* **69**, 22-32.

1912 Vasuki, Y., Holden, E. -J., Kovesi, P., Micklethwaite, S. (2017). An interactive image segmentation
1913 method for lithological boundary detection: A rapid mapping tool for geologists. *Computers and*
1914 *Geosciences* **100**, 27-40.

1915 Verwey, E. J. W. (1939). Electronic conduction of magnetite (Fe₃O₄) and its transition point at low
1916 temperatures. *Nature* **144**, 327-328.

1917 Vigh, D., Starr, B., Kapoor, J., Li, H. (2010) 3D full waveform inversion on a GOM data set. *80th*
1918 *Annual International Meeting, SEG, Expanded Abstracts*.

1919 Vigneresse, J. L. (1995). Crustal regime of deformation and ascent of granitic magma.
1920 *Tectonophysics* **249**, 187-202.

1921 Vigneresse, J. L., Tikoff, B. & Améglio, L. (1999). Modification of the regional stress field by
1922 magma intrusion and formation of tabular granitic plutons. *Tectonophysics* **302**, 203-224.

1923 Voigt, W. & Kinoshita, S. (1907). Bestimmung absoluter Werte von Magnetisierungszahlen,
1924 insbesondere für Kristalle. *Annalen der Physik* **329**, 492-514.

1925 Vollgger, S. A. & Cruden, A. R. (2016). Mapping folds and fractures in basement and cover rocks
 1926 using UAV photogrammetry, Cape Liptrap and Cape Paterson, Victoria, Australia. *Journal of*
 1927 *Structural Geology* **85**, 168-187.

1928 Waite, G. P. & Moran, S. C. (2009). VP Structure of Mount St. Helens, Washington, USA, imaged
 1929 with local earthquake tomography. *Journal of Volcanology and Geothermal Research* **182**, 113-122.

1930 Wall, M., Cartwright, J., Davies, R. & McGrandle, A. (2010). 3D seismic imaging of a Tertiary Dyke
 1931 Swarm in the Southern North Sea, UK. *Basin Research* **22**, 181-194.

1932 Wannamaker, P. E., Hasterok, D. P., Johnston, J. M., Stodt, J. A., Hall, D. B., Sodergren, T. L.,
 1933 Pellerin, L., Maris, V., Doerner, W. M., Groenewold, K. A., Unsworth, M. J. (2008). Lithospheric
 1934 dismemberment and magmatic processes of the Great Basin–Colorado Plateau transition, Utah,
 1935 implied from magnetotellurics. *Geochemistry, Geophysics, Geosystems* **9**, Q05019.

1936 Ward, K. M., Zandt, G., Beck, S. L., Christensen, D. H. & McFarlin, H. (2014). Seismic imaging of
 1937 the magmatic underpinnings beneath the Altiplano-Puna volcanic complex from the joint inversion of
 1938 surface wave dispersion and receiver functions. *Earth and Planetary Science Letters* **404**, 43-53.

1939 Warner, M., Ratclie, A., Nangoo, T., Morgan, J., Umpleby, A., Shah, N., Vinje, V., Stekl, I., Guasch,
 1940 L., Win, C., Conroy, G., Bertrand, A. (2013). Anisotropic 3D full-waveform inversion. *Geophysics*
 1941 **78**, R59-R80.

1942 Wauthier, C., Roman, D. C., Poland, M. P. (2013). Moderate-magnitude earthquakes induced by
 1943 magma reservoir inflation at Kīlauea Volcano, Hawaii. *Geophysical Research Letters* **40**(20), 5366-
 1944 5370.

1945 Wawrzyniak, P., Zlotnicki, J., Sailhac, P. Marquis, G. (2017). Resistivity variations related to the
 1946 large March 9, 1998 eruption at La Fournaise volcano inferred by continuous MT monitoring.
 1947 *Journal of Volcanology and Geothermal Research* **347**, 185-206.

1948 Weaver, J. (1994). *Mathematical methods for geo-electromagnetic induction*. Research Studies
 1949 Press.

- 1950 Westoby, M. J., Brasington, J., Glasser, N. F., Hambrey, M. J., Reynolds, J. M. (2012). “Structure-
1951 from-Motion” photogrammetry: A low-cost, effective tool for geoscience applications.
1952 *Geomorphology* **179**, 300-314.
- 1953 Whaler, K. A. & Hautot, S. (2006). The electrical resistivity structure of the crust beneath the
1954 northern Main Ethiopian Rift. In: Yirgu, G., Ebinger, C. J., Maguire, P. K. H. (eds) The Afar
1955 Volcanic Province within the East African Rift System. *Geological Society, London, Special*
1956 *Publications* **259**, 293-305.
- 1957 White, R. & McCausland, W. (2016). Volcano-tectonic earthquakes: A new tool for estimating
1958 intrusive volumes and forecasting eruptions. *Journal of Volcanology and Geothermal Research*
1959 **309**, 139-155.
- 1960 Wicks, C.W., Thatcher, W., Dzurisin, D., Svarc, J. (2006). Uplift, thermal unrest and magma
1961 intrusion at Yellowstone caldera. *Nature* **440**(7080), 72-75.
- 1962 Yamasaki, T., Kobayashi, T., Wright, T. J., Fukahata, Y. (2018). Viscoelastic crustal deformation by
1963 magmatic intrusion: A case study in the Kutcharo caldera, eastern Hokkaido, Japan. *Journal of*
1964 *Volcanology and Geothermal Research* **349**, 128-145.
- 1965 Zandomenighi, D., Aster, R., Kyle, P., Barclay, A., Chaput, J., Knox, H. (2013). Internal structure
1966 of Erebus volcano, Antarctica imaged by high-resolution active-source seismic tomography and
1967 coda interferometry. *Journal of Geophysical Research-Solid Earth* **118**, 1067–1078.

1968

1969 **7. Figure captions**

1970

- 1971 Figure 1: Schematic of a vertically extensive, transcrustal magma plumbing system involving
1972 transient, interconnected, relatively low-volume tabular magma intrusions (e.g., dykes, sills, and
1973 laccoliths) within a crystal mush (based on Cashman *et al.*, 2017; Cruden *et al.*, 2018).

1974

1975 Figure 2: (A) Interferograms showing fringes caused by the pressurisation of a point source directly
1976 beneath a stratovolcano from both ascending and descending satellite lines of sight. Note that the
1977 centre of the fringes are slightly offset from the summit of the volcano (marked by a black triangle).
1978 (B) Typical fringe patterns for analytical deformation sources in an elastic half space from
1979 ascending satellite geometry: (i) Mogi source at 5 km depth; (ii) dyke extending between depths of
1980 3 and 9 km; (iii) rectangular sill; and (iv) a penny-shaped horizontal crack both at 5 km depth.

1981

1982 Figure 3: (A) Ascending line of sight (LOS) co-eruptive interferogram from the 2008 basalt lava
1983 extrusion between the Alu and Alu South domes and the Dalafilla stratovolcano (modified from
1984 Pagli *et al.*, 2012). (B) Inversion of uplift and subsidence patterns, recorded by InSAR during the
1985 2008 basalt lava eruption at the Alu dome in the Danakil Depression, suggested ground deformation
1986 could be attributed to a combination of: (i) deflation of a reservoir, modelled as a Mogi source, at
1987 ~4 km depth; (ii) inflation and deflation of a tabular sill at ~1 km depth; and (iii) opening of a dyke
1988 beneath the eruptive fissure (Figs 3A and B) (Pagli *et al.*, 2012). See Figure 3A for location. (C)
1989 Geological map showing that lava flows radiate out from Alu and originate from the periphery of
1990 the dome, which is cross-cut by an array of randomly oriented faults (modified from Magee *et al.*,
1991 2017). (D) Magee *et al.*, (2017) inferred Alu is underlain by a saucer-shaped sill plumbing system,
1992 based on field observations and comparison to seismic reflection data, not a tabular sill (Fig. 3B).

1993

1994 Figure 4: Example of integrating seismology and petrology to constrain time-scales of magma
1995 storage and recharge (from Saunders *et al.*, 2012). Calculated Fe-Mg diffusion time scales of
1996 orthopyroxene crystals compared to monitoring data for the same eruptive period for Mount St.
1997 Helens. (A) The seismic record of depth against time of the 1980–1986 eruption sequence. (B)
1998 Measured flux of SO₂ gas. (C) Calculated age of orthopyroxene rim growth binned by month for the
1999 entire population. The age recorded is the month in which the orthopyroxene rim growth was
2000 triggered by magmatic perturbation. The black line displays the running average (over five points,

equivalent to the average calculated uncertainty in calculated time scales) of all the data. The peaks in the diffusion time series correspond to episodes of deep seismicity in 1980 and 1982 and to elevated SO₂ flux in 1980 and possibly 1982. (D) Running average of the orthopyroxene rim time scales, displaying reverse zonation (Mg-rich rims) in blue and normal zonation (Fe-rich rims) in green. There are reverse zonation peaks in the early 1980, probably due to rejuvenation of the magma system by hotter pulses, whereas Fe-rich rims are more dominant from 1982 on. Vertical dashed grey lines represent the volcanic eruptions.

Figure 5: Plot of melt inclusion saturation and earthquake hypocentre depths, which suggest magma storage occurred at 1–5 km depths, beneath the Dabbahu volcanic system in Afar, Ethiopia (modified from Field *et al.*, 2012). Melt inclusion data obtained from analyses of alkali feldspar, clinopyroxene, and olivine phenocrysts within Dabbahu lavas <8 Kyr (Field *et al.*, 2012). Earthquake data recorded during the 2005 dyke event (Ebinger *et al.*, 2008).

Figure 6: (A) P-wave (V_p) tomography beneath Montserrat (black outline), highlighting the location of fast and slow seismic velocity anomalies (i.e. >6% faster or slower than average) relative to the location of the Silver Hills (SH), Central Hills (CH), and Soufrière Hills (SHV) volcanoes (modified from Shalev *et al.*, 2010). The fast velocity anomalies, interpreted to represent solidified andesitic intrusions underlie the volcanoes (Shalev *et al.*, 2010).

Figure 7: Static and dynamic gravimetric investigations of two active silicic magmatic systems in the Andes: Uturuncu volcano (Bolivia; A, C, and E) and the Laguna del Maule volcanic field (Chile; B, D, and F). (A) 3D view of the isosurface corresponding to the -120 kg m³ density contrast beneath Uturuncu volcano, derived from Bouguer gravity data, interpreted to reflect a large (~750 km³) plumbing system composed of a lower (<10 km) partially molten reservoir and upper, fractured and fluid-bearing solidified intrusions above sea level (after del Potro *et al.*, 2013). (B) 3D

view of the -600 kg m^3 density contrast isosurface beneath the Laguna del Maule, which is interpreted to define a magma reservoir ($>50 \%$ melt) within a larger region of a crystal mush system; the 2D planes show slices through the dataset (Miller *et al.*, 2017). Elevation above sea level (a.s.l.) shown. See Figure 7D for area of data coverage. (C) Map of the 55 km long, dynamic gravity network (white circles) installed to track changes in gravity over time and space at Uturuncu volcano between 2010 and 2013 (modified from Gottsmann *et al.*, 2017). (D) Spatio-temporal residual gravity changes at Laguna del Maule recorded from 2013–2014, after correcting for deformation effects (modified from Miller *et al.*, 2017). (E) Gravity and deformation data, recorded from Uturuncu from 2010–2013, plotted against the measured free-air gravity gradient (solid red line) and associated errors (broken red lines) (modified from Gottsmann *et al.*, 2017). The data follow the gradient and are indicative of a subsurface density change as a cause of the uplift, possibly reflecting the release of fluids from a large deep-seated magma reservoir (i.e. the Altiplano-Puna Magmatic Body; Chmielowski *et al.*, 1999) through the vertically extensive crystal mush system shown in (A) (Gottsmann *et al.*, 2017). (F) Plot of gravity against horizontal distance for the source centre at Laguna del Maule (modified from Miller *et al.*, 2017). The increase in gravity of up to $120 \mu\text{Gal}$ is explained by a hydrothermal fluid injection focused along a fault system, shown in (D), at 1.5–2 km depth as a result of a deeper seated magma injection, and is best modelled by a vertical rectangular prism source.

Figure 8: Gravity changes and deformation at the restless Long Valley caldera. (A) Map of the Long Valley caldera, California, USA, which hosts a resurgent dome (black outline), to highlight changes in residual gravity between 1982 and 1999 (modified from Tizzani *et al.*, 2009). (B) Plot of ground uplift and residual gravity changes with radial distance from the centre of the resurgent dome in (A) (modified from Tizzani *et al.*, 2009). The correlation between uplift and positive gravity residuals across the resurgent dome indicates ground deformation was instigated by intrusion of magma (Tizzani *et al.*, 2009).

2053

2054 Figure 9: (A) Map showing MT stations deployed around Volcán Uturuncu (U) and Volcán
2055 Quetena (Q), relative to areas of uplift and subsidence (modified from Comeau *et al.*, 2015). The
2056 white box shows area of modelled 3D MT data (Comeau *et al.*, 2015). (B) Regional 2D
2057 magnetotelluric line through the Altiplano-Puna magma body (APMB) highlighting the position of
2058 Volcán Uturuncu (modified from Comeau *et al.*, 2015). The APMB corresponds to a large,
2059 conductive (i.e. low-resistivity) body (Comeau *et al.*, 2015; Comeau *et al.*, 2016). Above the APMB
2060 are other areas of low-resistivity (e.g., C4) that are likely upper crustal magma reservoirs and dykes
2061 (Comeau *et al.*, 2016). C1–C7 and R1–R2 identify discrete zones of marked conductivity or
2062 resistivity, respectively (see Comeau *et al.*, 2015; Comeau *et al.*, 2016 for details). The white box
2063 shows area of modelled 3D MT data (Comeau *et al.*, 2015). See Figure 9A for location.

2064

2065 Figure 10: (A) Interpreted seismic section and geological map showing the distribution of and
2066 connectivity between sills within the Faroe-Shetland Basin (modified from Schofield *et al.*, 2017).
2067 Mapping of magma flow patterns within individual sills reveals that the sill-complex facilitates
2068 extensive vertical and lateral magma transport. Magma was fed into the sedimentary basin via
2069 basement-involved faults. TWT = two-way travel time. (B) Interpreted seismic section and
2070 geological map describing the spatial relationship between volcanoes/vents and sills, inferred to
2071 represent the magma plumbing system, emplaced at ~42 Ma (modified from Jackson *et al.*, 2013;
2072 Magee *et al.*, 2013a). Sills are laterally offset from the volcanoes/vents summits. No ‘magma
2073 chambers’ are observed in the seismic data, which images down to ~8 s TWT (i.e. ~>10 km)
2074 (Magee *et al.*, 2013a).

2075

2076 Figure 11: (A) Interpreted seismic section from the Exmouth Sub-basin offshore NW Australia,
2077 which images a saucer-shaped sill that is overlain by a forced fold and feeds a small vent from its
2078 inclined limb (modified from Magee *et al.*, 2013b). See Figure 11B for line location). (B) Time-

2079 structure map of the folded horizon (thick black line) in (A), highlighting fault traces and vent
2080 locations and thicknesses (modified from Magee *et al.*, 2013b). (C) Seismic section from the
2081 Farsund Basin, offshore southern Norway, which images part of a dyke-swarm that has been rotated
2082 by basin flexure post-emplacement (modified from Phillips *et al.*, 2017).

2083
2084 Figure 12: (A) At the sample scale, all magnetic grains create a magnetic fabric. (i) Dominantly
2085 prolate fabric, where K_2 and K_3 are least certain and form a girdle. Only the magnetic lineation (K_1)
2086 can be confidently determined. (ii) When $K_1 > K_2 > K_3$, both a foliation (K_1 – K_2) and a lineation (K_1)
2087 may be discerned, defining a triaxial fabric. (iii) When K_1 and K_2 are equally uncertain and form a
2088 girdle, K_3 is perpendicular to a foliation. (B) Schematic representation of how magma flow within a
2089 planar sheet intrusion can produce imbricated magnetic fabrics at its margins, the closure of which
2090 define the magma flow direction (after Féménias *et al.*, 2004). (C) AMS data and interpretations
2091 from part of the Trawenagh Bay Granite, NW Ireland (adapted from Stevenson *et al.*, 2007a). (i)
2092 AMS foliation traces are shown in blue and lineation traces in red. Lobes were defined in this
2093 intrusion based on foliations curving around a lineation axis. In some lobes, the magnetic lineation
2094 trend was parallel to this axis, whilst in others they tended to splay or converge down flow. (ii) 3D
2095 sketch showing the geometry of three of the lobes (numbered in part i).

2096
2097 Figure 13: (A) Starting model derived from smoothed, pre-stack, time-migrated (PSTM) stacking
2098 velocities. (B) Final 2D FWI-derived velocity model obtained using 10 km streamer data and
2099 inversion frequencies of between 2.5 and 24 Hz. (C) FWI velocity model overlain by the 2D pre-
2100 stack, depth-migrated (PSDM) section. Strong irregular reflections in the lower half of the section
2101 are from basaltic intrusions, which appear as high-velocity anomalies in the FWI velocity model.
2102 Both the FWI velocity model and the PSDM pick out a major unconformity, and show shallow
2103 channels in the upper parts of the section (redrawn from Kalincheva *et al.*, 2017).

2105 Figure 14: (A) UAV orthophotograph of the wave cut platform at Bingie Point, NSW, Australia
2106 showing the distribution of Palaeogene dolerite (Dol) and dacite (Dac) dykes within Devonian
2107 tonalite (Ton), diorite (Di), and aplite (Ap) host rocks. (B) Circular histogram of joint sets measured
2108 in the Devonian rocks from the orthophotograph; the dominant (purple) set is parallel to and likely
2109 contemporaneous with the Palaeogene dykes. (C) Annotated close-up image highlighting dykes and
2110 structural features. The northern dacite dyke shows two broken bridge (BB) structures, whilst the
2111 central dolerite dyke displays prominent step structures (S). Narrow apophyses are also associated
2112 with the broken bridges and steps.

Figure 1

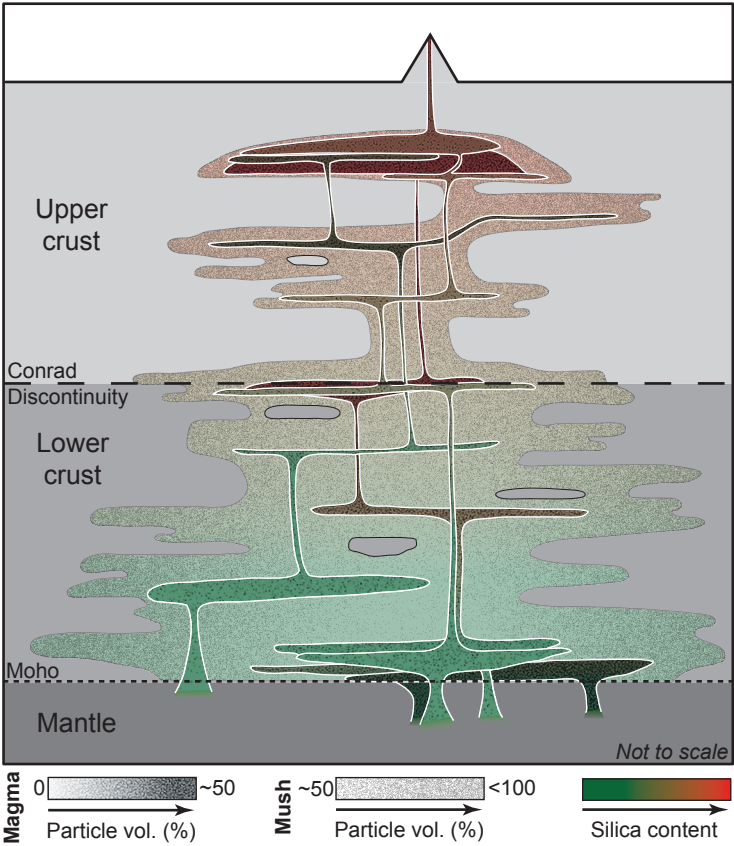


Figure 2

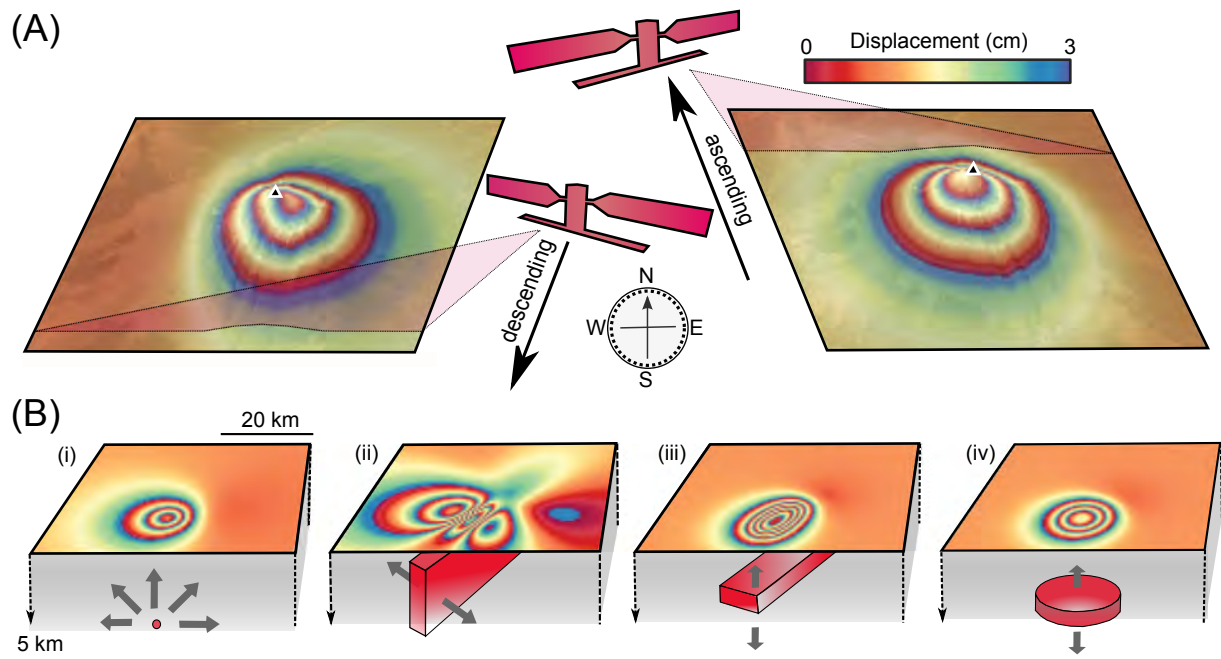


Figure 3

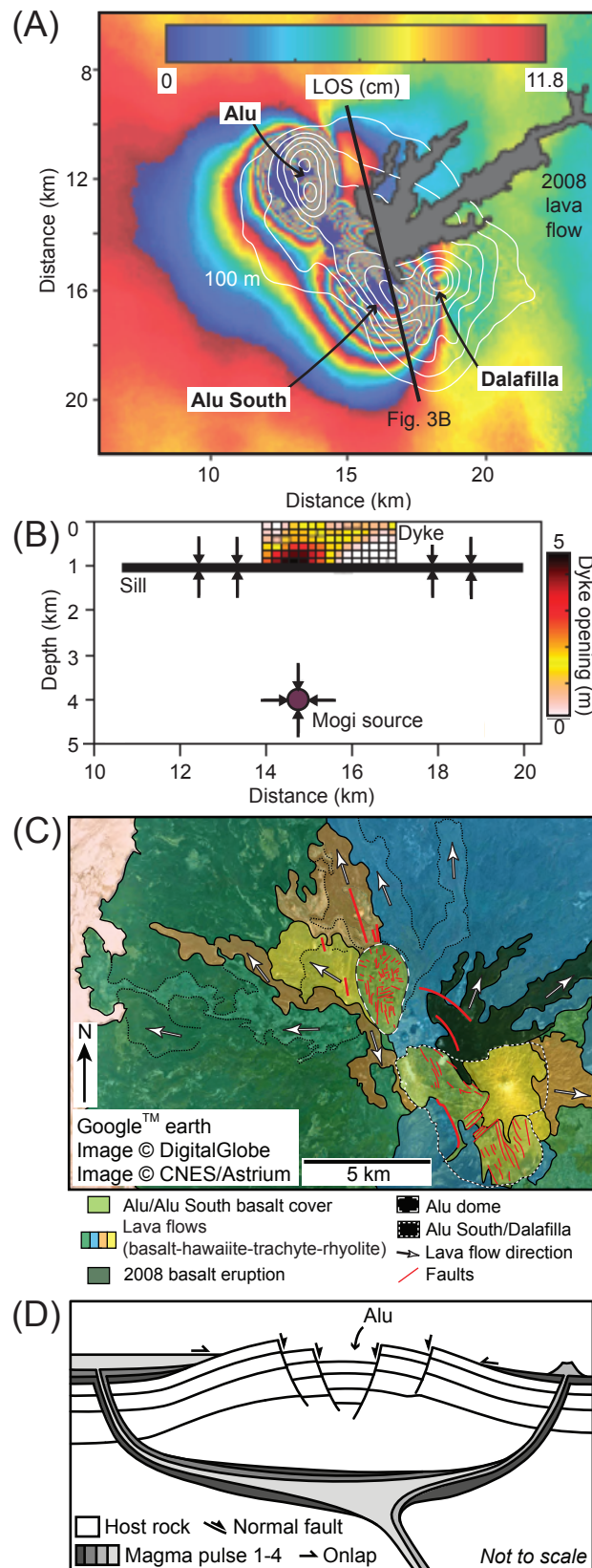


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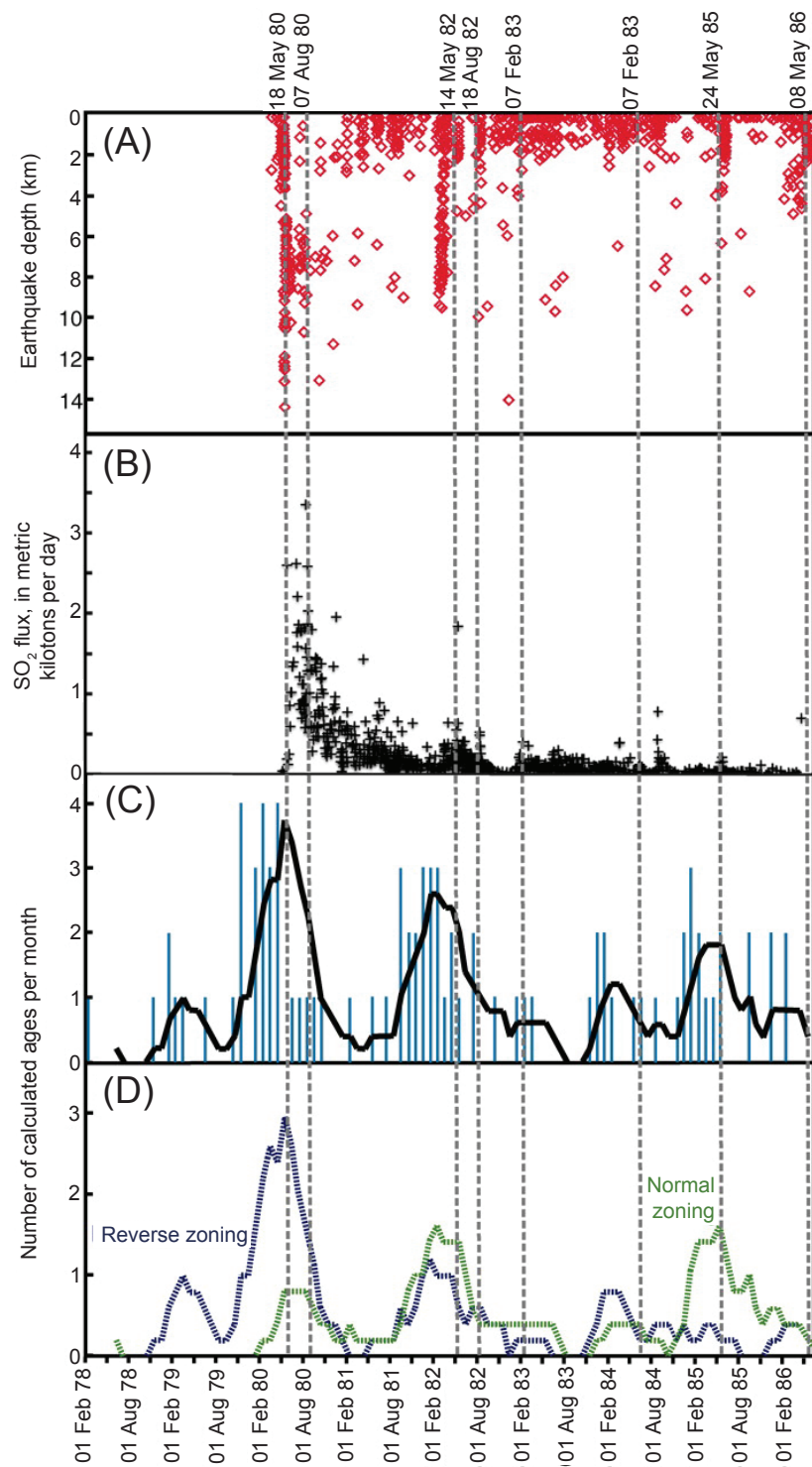


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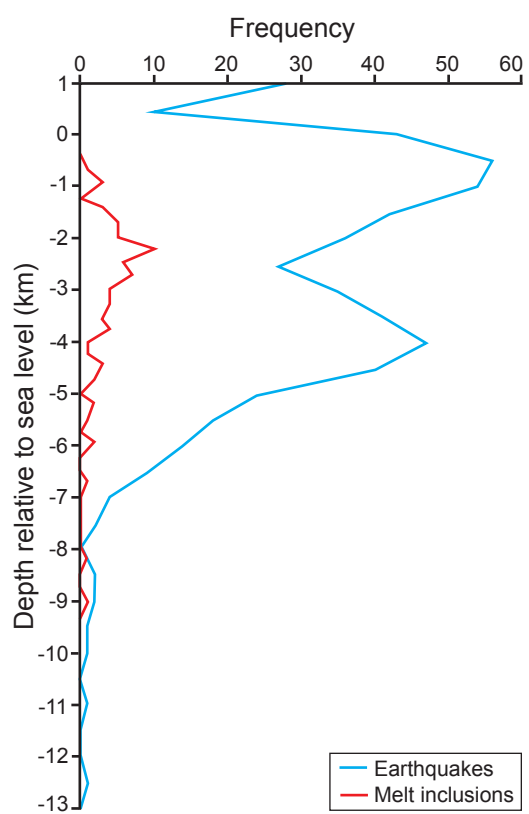


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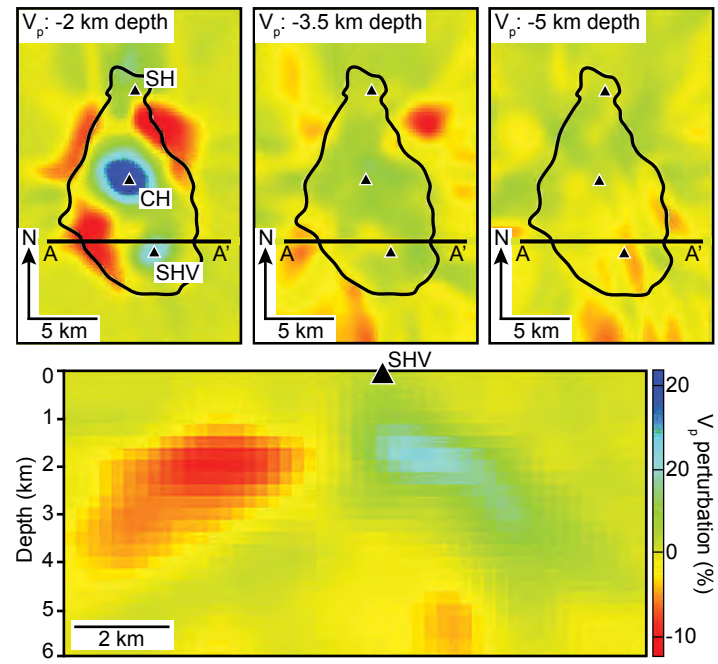
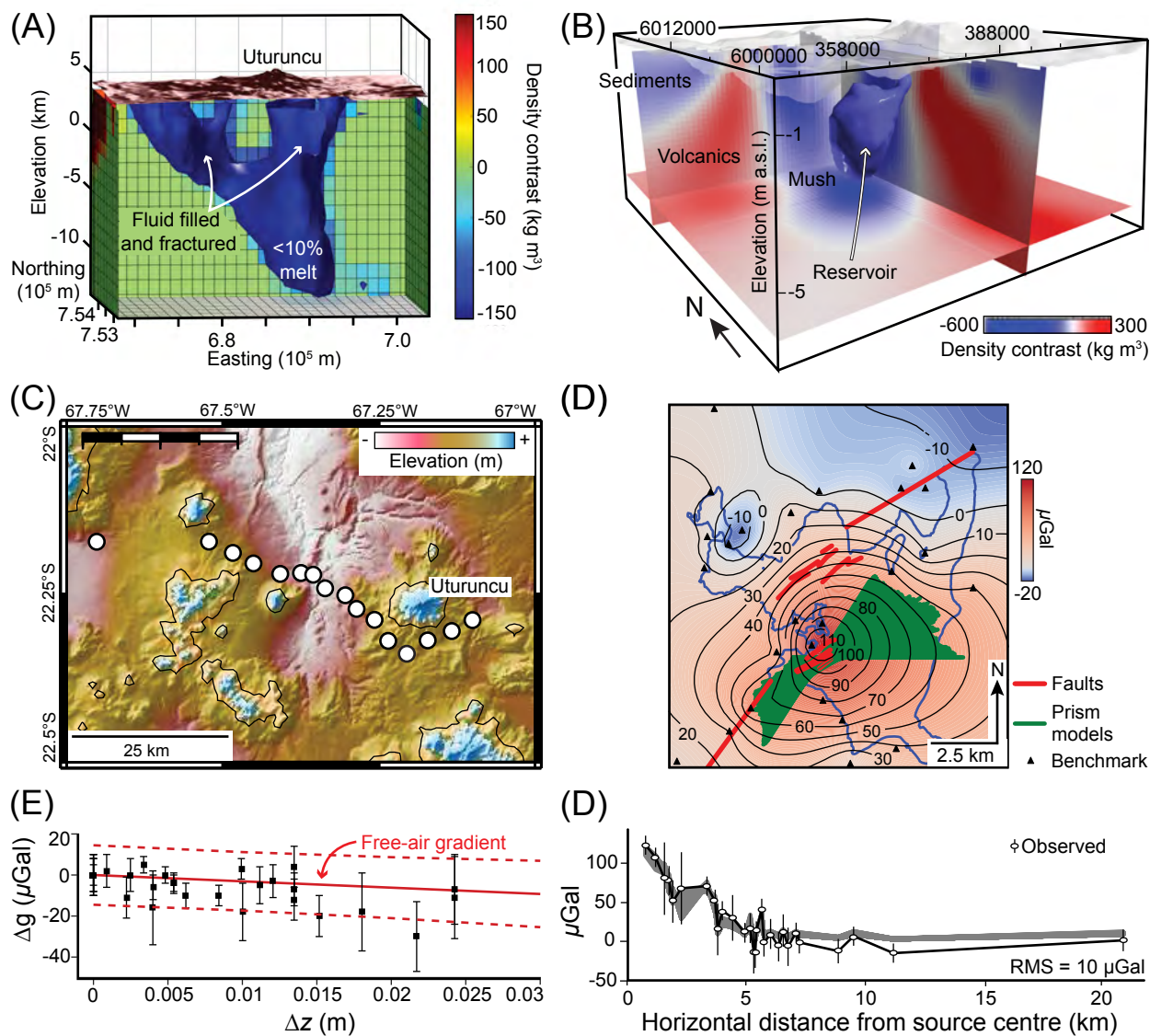


Figure 7



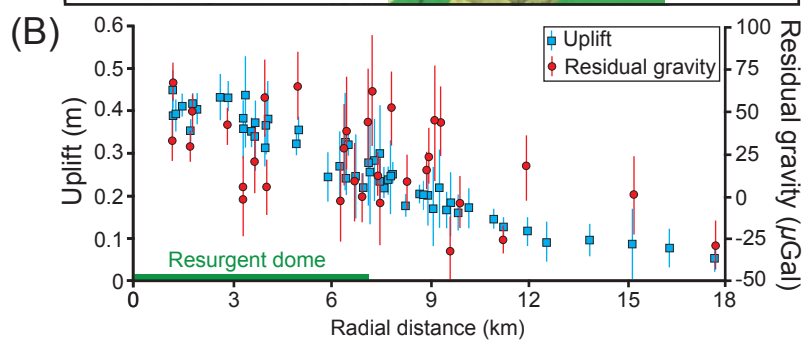
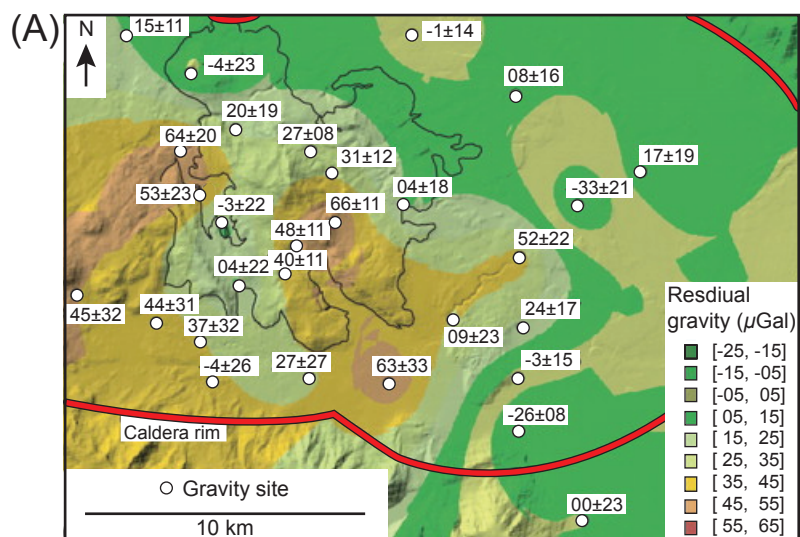


Figure 9

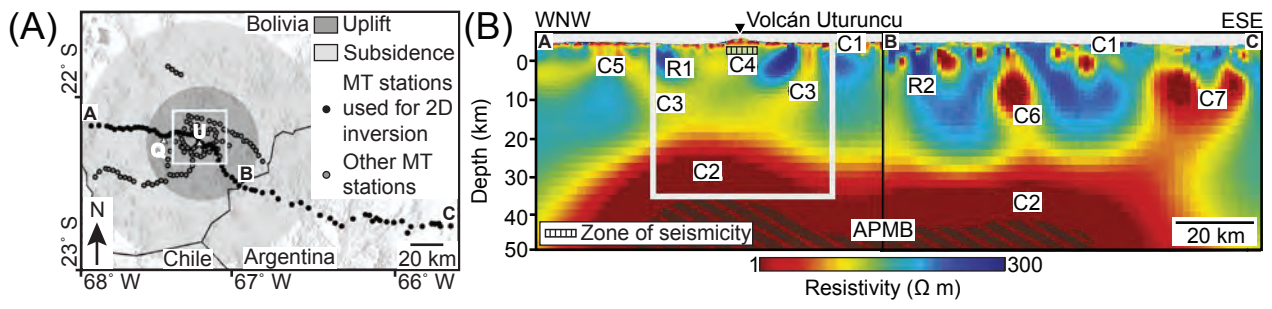


Figure 9

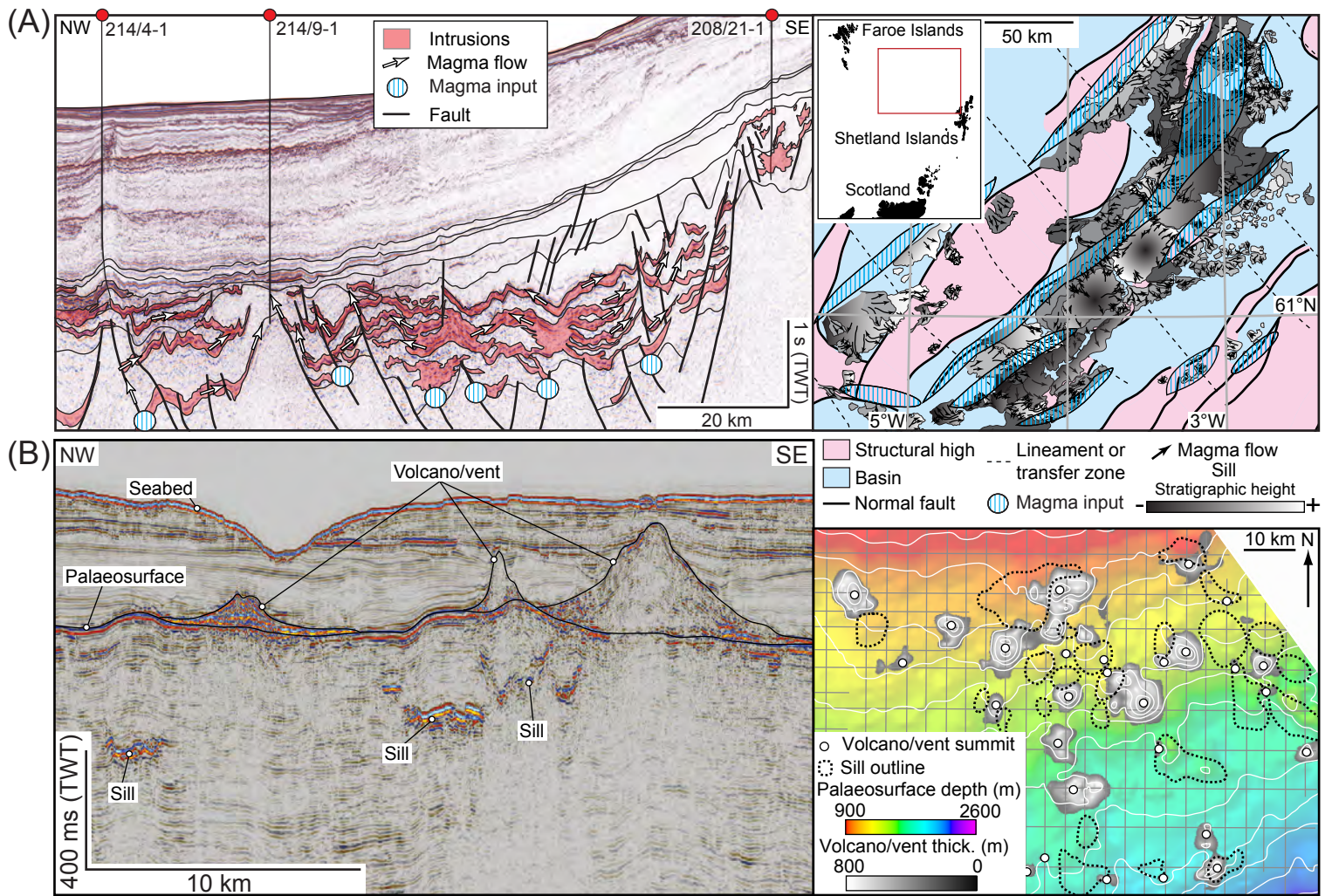


Figure 10

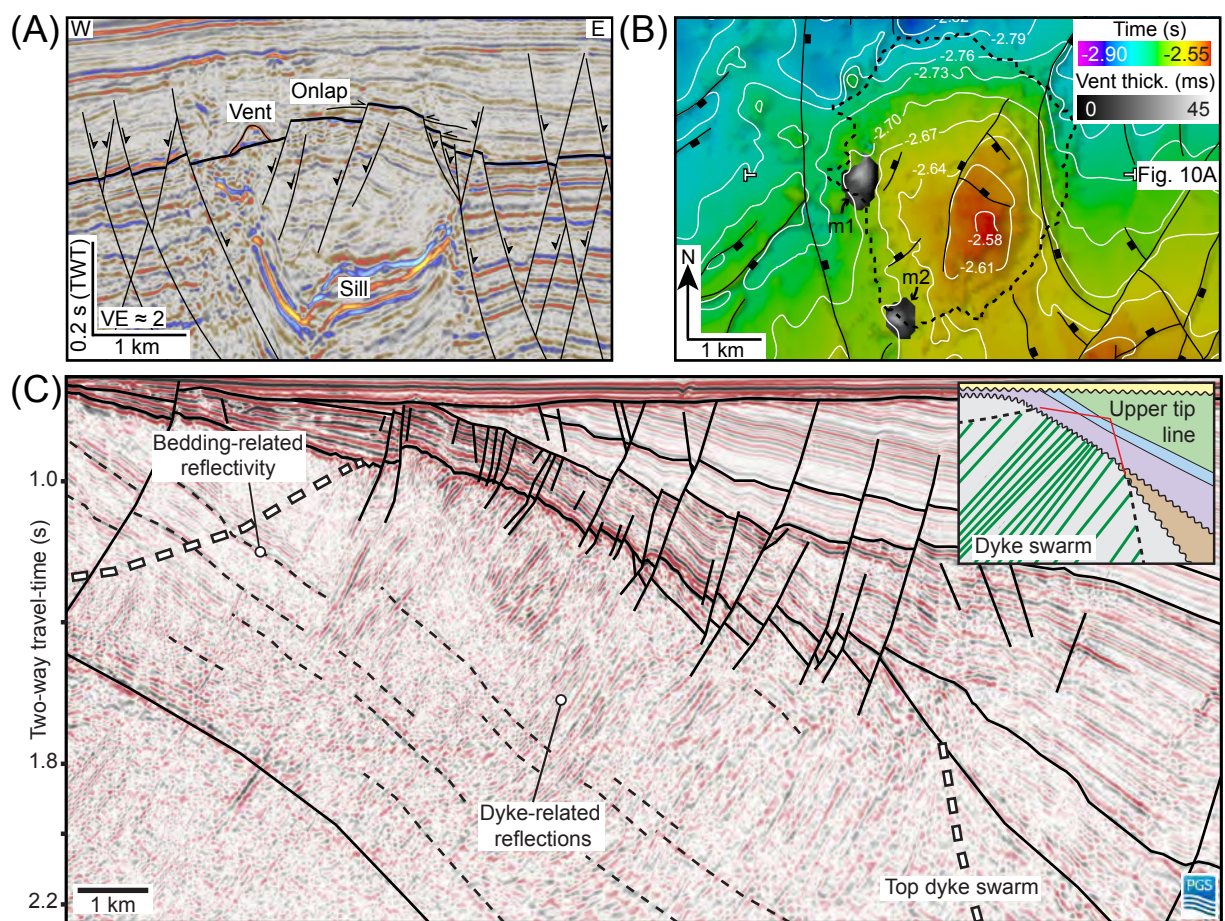


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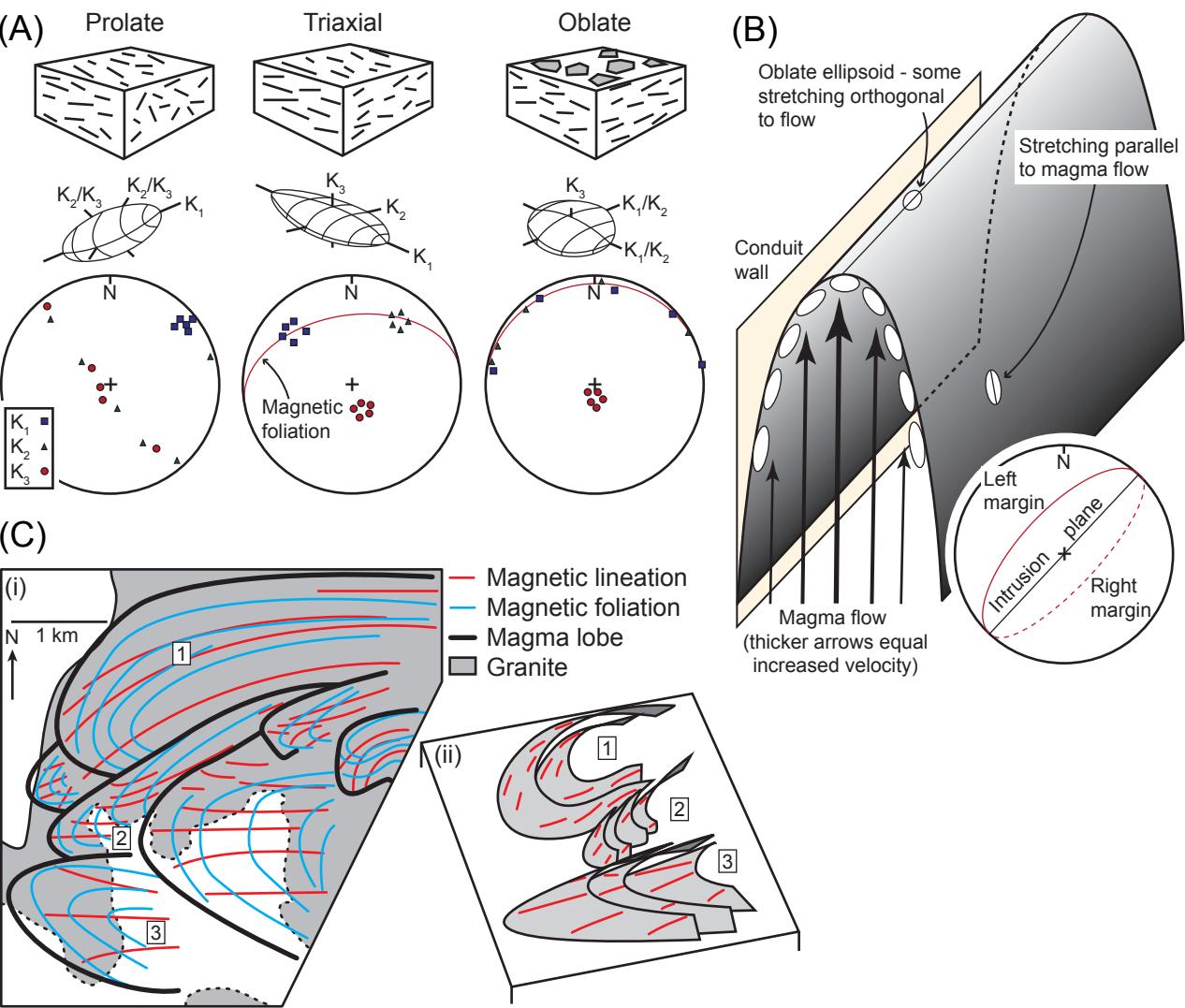


Figure 12

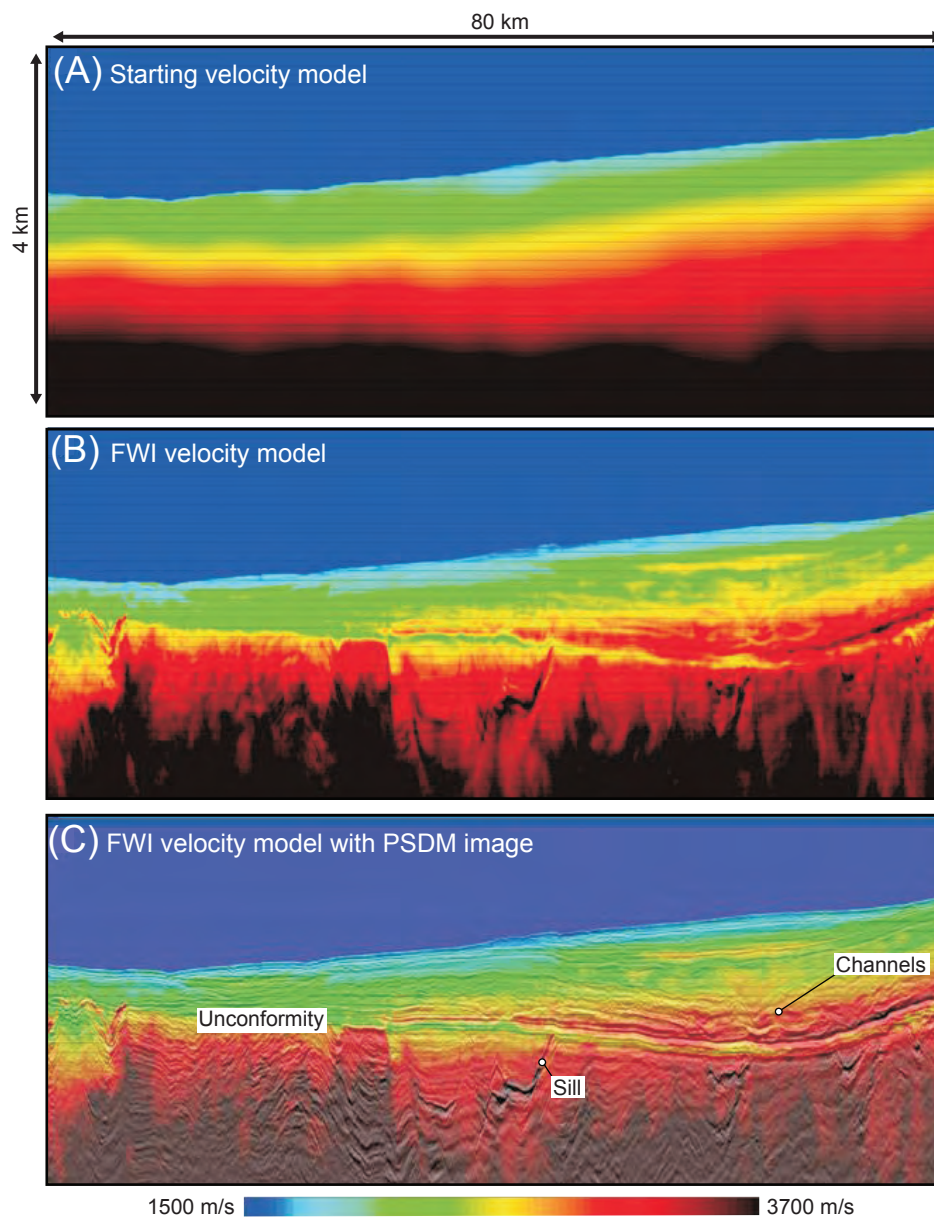


Figure 13

